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OBSERVATION AND PREDICTION OF THE 1986 BREAKUP ON THE ATHABASCA RIVER UPSTREAM OF FORT MCMURRAY, ALBERTA

by

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ABSTRACT

Breakup on the Athabasca River between the mouth of the Pembina River and Fort McMurray took place on April 19, after 62°C-days of accumulated melt over a period of about 26 days. Prior to breakup, the ice cover deterioration was characterized by a gradual increase in the fraction of open water from 0.02 on April 3 to 0.38 on April 18. In general, the breakup could be characterized primarily as thermal, and perhaps overmature, because very low stage increases (on the average) were apparent, most probably due to the below average snowfall.

A numerical model, developed to predict the generation of open water, was tested against one year of observations and was found to function very well. Using measured values of daily air temperature, solar radiation and discharge and initializing the model by defining the maximum winter ice thickness, the day of the first consistent stage increase and the last day of accumulated snow on the ice cover, it was found that the observations of open water could be calculated using an exposure factor of 0.75, an albedo of ice and water of 0.3 and 0.05, respectively, and a convection heat transfer coefficient over ice and water of 10 W/m^2-°C and 15 W/m^2-°C, respectively.

Sensitivity analysis suggested that the variations of the albedos and the heat transfer coefficients within appropriate and accepted limits did not significantly alter the model performance. It was most important to have a good definition of the maximum winter ice thickness, the date of first thinning (last day of a substantial snow cover) and the first day of appreciable rise in stage.

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P. Mostert of the Resource Technologies Department, Alberta Research Council, performed most of the field work in documenting the breakup progression. Simone Oatway and Carolyn Reitzel typed the final manuscript. M. Quazi reviewed the final draft of the report.

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1.0 INTRODUCTION

1.1 Background

The City of Fort McMurray, Alberta, is located on the Athabasca River at the confluence of the Clearwater River. Ice jams, which have been documented as far in the past as 1875 (Blench, 1963), often form downstream of the confluence and cause serious flooding within the lower portion of the city. The most recent of these events occurred in 1979 (Doyle and Andres, 1979) and since then considerable effort has been spent observing breakup upstream of Fort McMurray and characterizing the resulting ice jams.

Although the ice jams that occur on an annual basis have been well characterized and ice jam rating curves have been developed (Andres and Doyle, 1984) to relate ice jam water levels to the discharge, very little is known about the breakup process itself. This deficiency, despite a number of years of observations (Andres and Rickert; 1984,1985) arises from the very dynamics of the process, the extremely long length of the river reach which must be observed and the general inaccessibility of the area. Nevertheless, to mitigate potential ice jam related flood damages, an operational system is necessary to forecast the timing of the onset of breakup, hereby shortening the window of concern during the March and April melt period.

This understanding of breakup is being improved by the observation of the breakup process. A somewhat rational view of how the snowpack levels and temperatures affect the progression of breakup from small tributaries down to the lower Athabasca River is developing. The observations of the 1986 breakup are particularly useful. The systematic measurement of areas of open water on the Athabasca River, over a period of about three weeks, allows for a quantitative assessment of the melt and breakup processes which can then be related to atmospheric conditions.

1.2 Objectives

The long term objectives of the ice observation program on the Athabasca River are to gain an understanding of the physical breakup processes. This is to lead towards the development of forecasting techniques to predict the onset and severity of breakup, so that flood damages can be both predicted and managed.

This report describes the 1986 breakup observations on the Athabasca River upstream of Fort McMurray (figure 1). It also addresses a simple model which may be applicable in forecasting the onset of breakup. Included in the report will be a brief literature review of breakup prediction, a summary of the meteorological conditions (air temperature and solar radiation) prevailing in the Athabasca River basin during the 1986 spring melt period, and estimates of the depletion of the snowpack and resulting discharge increases in the Athabasca River. The applicability of the model will be assessed, a calibration carried out and a sensitivity analysis will be done to identify the most relevant model parameters. Conclusions and recommendations will also be made with respect to the direction and extent of further work.

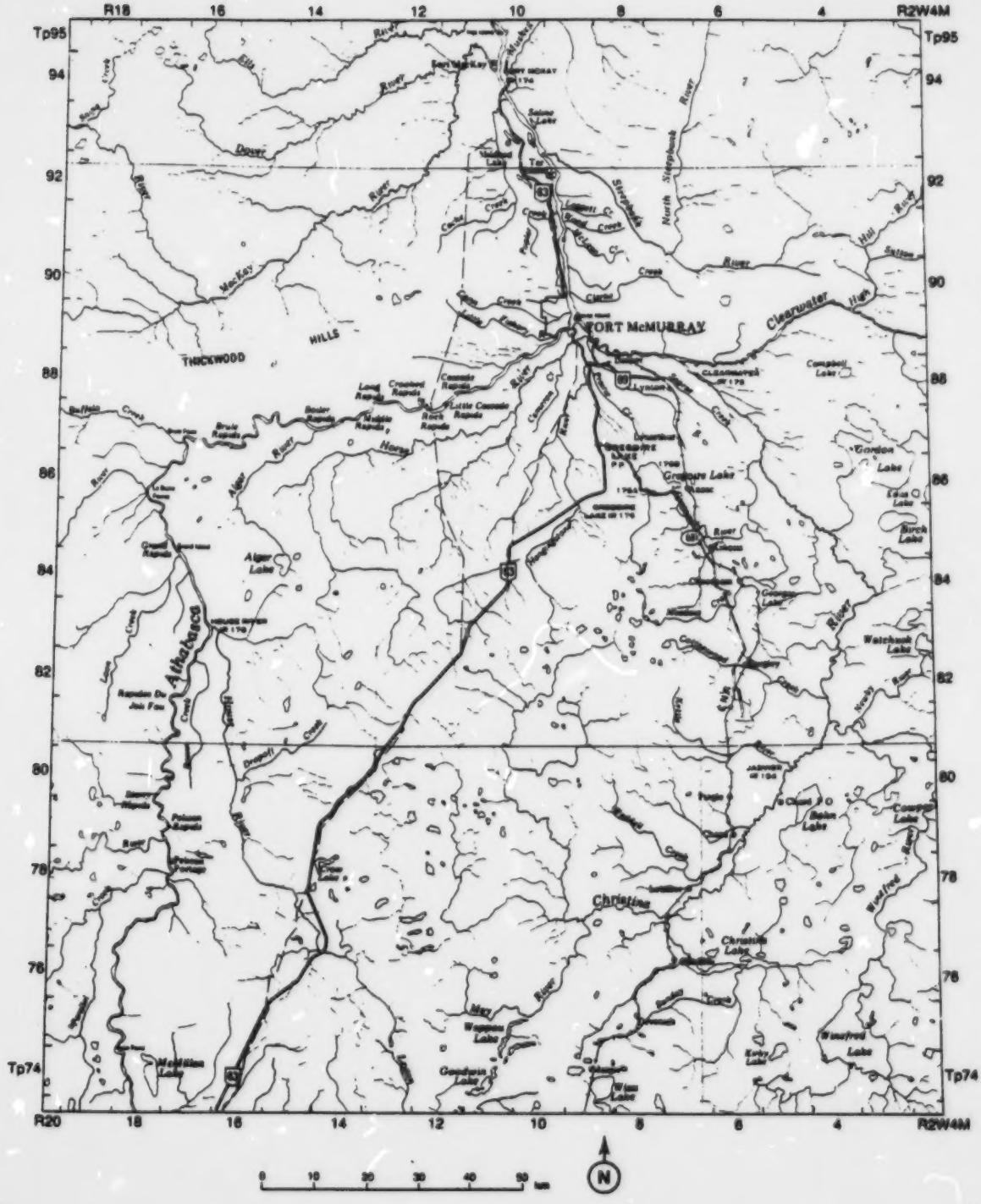


Figure 1 Location plan

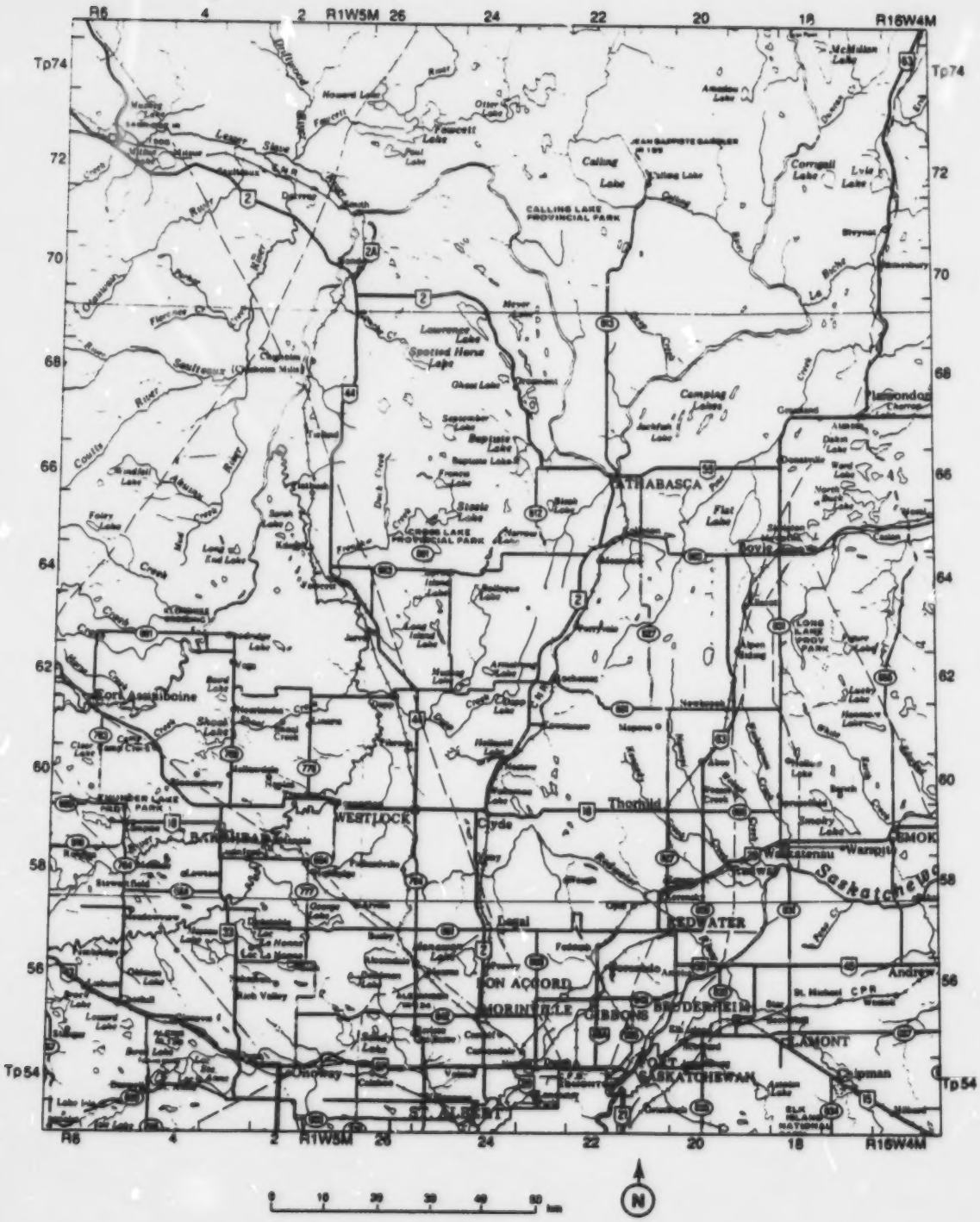


Figure 1 (cont'd) Location plan

2.0 METEOROLOGIC CONDITIONS

2.1 Winter Snowpack and Snowmelt

The winter snowpack was characterized using Alberta Environment snow course data and the Twin Lakes snow pillow information. In the western, southern and central plains areas of the Athabasca River basin, the 1985/86 winter snowfall averaged about 60% of the twelve year mean prior to 1985 (Alberta Environment, 1986). At the end of February 1986, the accumulated water equivalent was about 40 mm, on average, or about 57% of normal. This increased slightly to 44 mm or 62% of normal during the third week of March, just prior to the beginning of the melt period. By the first week of April, the water equivalent of the snow on the ground, was reduced to 20 mm or about 30% of normal. At this late date, there was extreme variability in the amount of snow in the basin, and many of the snow courses had no snow. It appears that the winter of 1985/86 was below average in terms of snow accumulation. It is also apparent that snowmelt was completed somewhat earlier than other years, resulting in the relatively small amount of snow available for melt in early April.

Comparison of the snow course data with the Twin Lakes snow pillow information suggests that there is a reasonable correlation between that one point and the basin average (figure 2). The Twin Lakes water equivalent in early March is the same as the basin average. However, in the third week of March, Twin Lakes tends to overestimate the basin mean, whilst in the first week of April, it underestimates the basin mean. Nevertheless, it is apparent that the maximum snow accumulation occurred early in the last week of March, at which time significant and steady melt began, and there was a net depletion of the snowpack. The melting of the snowpack began on March 25 and progressed rapidly over a period of ten days, and melt was completed after about 40°C-days of accumulated temperature. It is interesting that the snowpack at the snow pillow was negligible on or about the same time as the first aerial reconnaissance was made and at which time open water leads were just beginning to make an appearance.

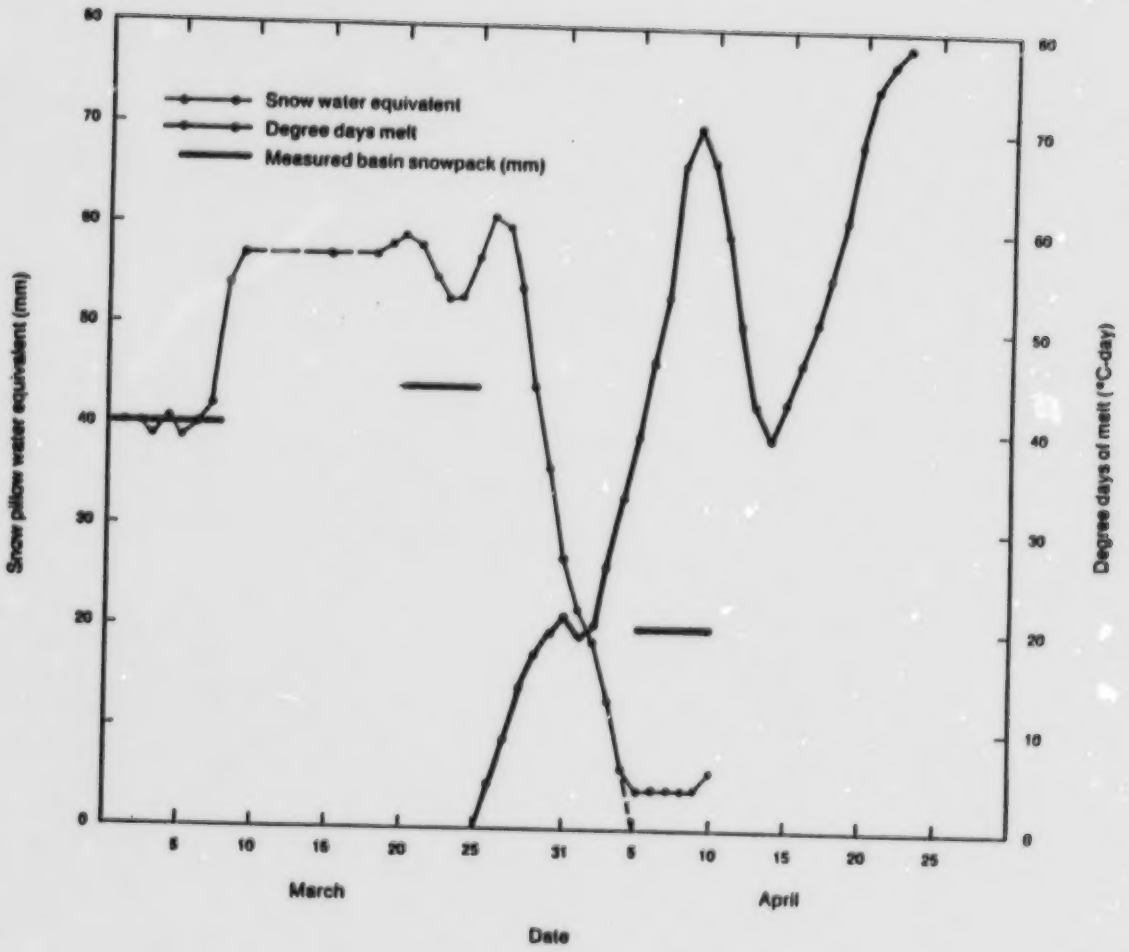


Figure 2 Degree days of melt and snow pack reduction at the Twin Lakes snow pillow, March and April 1986.

2.2 Air Temperature

The temperature regime of the Athabasca River basin was characterized from AES records at Whitecourt, Edmonton, Slave Lake and Fort McMurray. The mean basin air temperature was computed using the Thiesson polygon technique which produced the relationship

$$[1] \quad T_a = 0.39\ T_w + 0.06\ T_e + 0.36\ T_{sl} + 0.19\ T_{fm}$$

where T_a is the mean basin air temperature (either daily mean, maximum or minimum) and T_w, T_e, T_{sl} and T_{fm} are the air temperatures at Whitecourt, Edmonton, Slave Lake and Fort McMurray, respectively.

During the month of March, the maximum daily temperature seldom rose above 10°C and the minimum daily temperature never reached 0°C (figure 3). A brief warm spell occurred during the first week of April when the maximum temperature reached 24°C, but this was followed by a signifi- cant cooling trend for about six days when the maximum temperature never rose above 0°C. It is interesting that during the entire melt/breakup period, the minimum temperature exceeded 0°C on less than five days. This suggests that the melt was not vigorously driven by the air temperatures.

The first positive degree days (melting) occurred on March 25 (figure 2) and by April 9, about 70°C-days of melt had been accumulated. However, very cold conditions after April 9 drove down the accumulated heat index to almost half its peak value, before additional gains were evident on April 14. The maximum of 70°C-days was not re-established before the breakup was completed on April 20.

2.3 Solar Radiation

The average basin solar radiation was computed on the basis of measured solar radiation at Edmonton and measured hours of bright sunshine at Edmonton, Slave Lake and Fort McMurray. As for the air temperatures, the computed basin solar radiation was determined by applying the

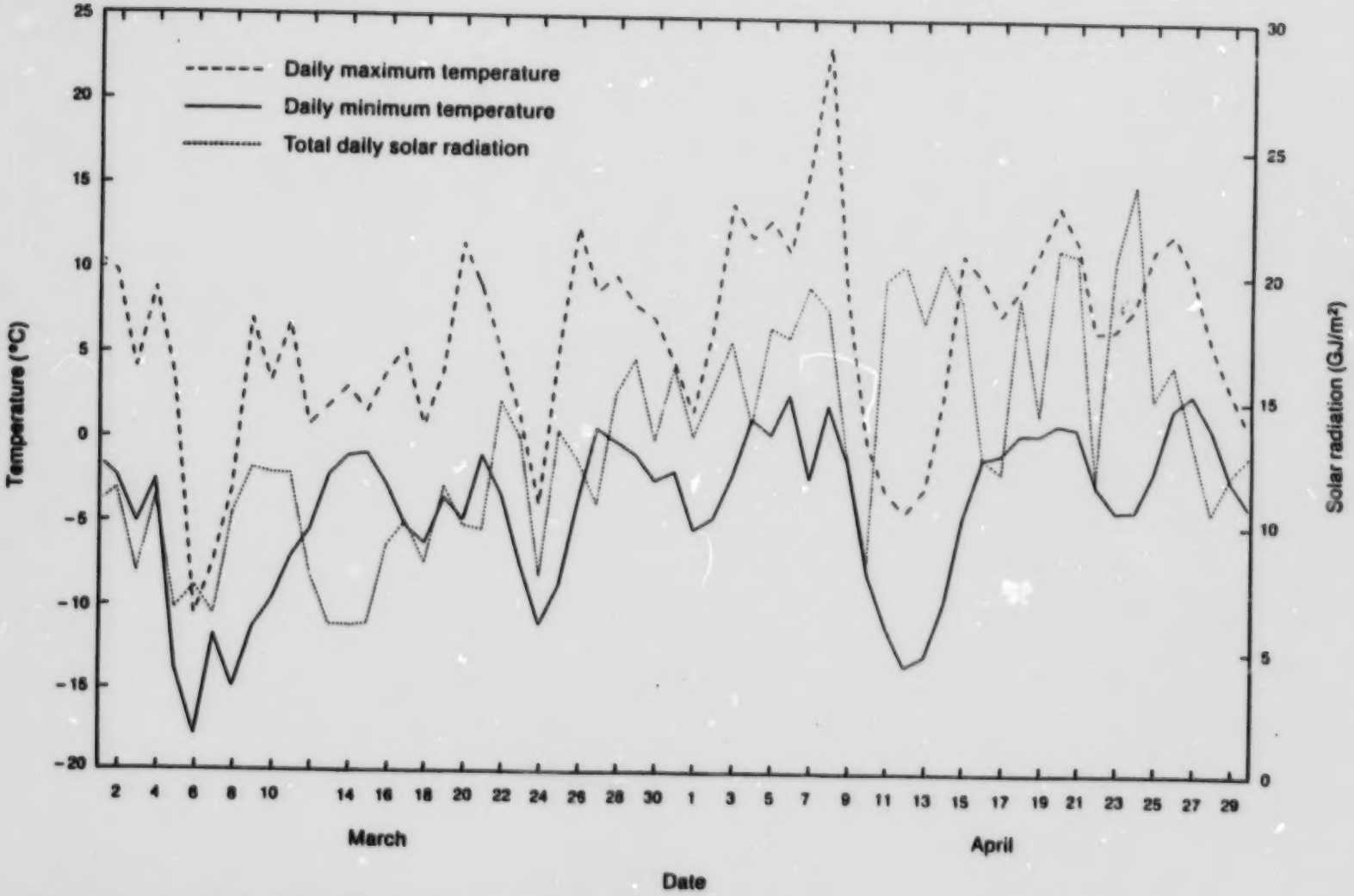


Figure 3 Daily maximum and minimum air temperatures and solar radiation over the Athabasca River basin, March and April 1986

Thiesson polygon technique to the three relevant meteorology stations. Thus, the mean basin daily solar radiation was computed by:

$$[2] \quad H = 0.29\ H_e + 0.52\ H_{sl} + 0.21\ H_{fm}$$

where H_e, H_{sl} and H_{fm} are the estimate daily total radiation received on a horizontal surface at Edmonton, Slave Lake and Fort McMurray, respectively.

2.3.1 Solar Radiation Estimates

Other than at Edmonton, the only consistent solar variable measured at most meteorologic stations in, or in the vicinity of, the Athabasca River basin is hours of bright sunshine. Solar radiation can be estimated at these stations by transforming hours of bright sunshine or percentage of maximum possible bright sunshine for any given day to daily estimates of solar radiation. This requires a relationship between the fraction of daily bright sunshine and the fraction of maximum possible solar radiation received on any given day.

The angle of the sun, α for each hour between sunrise, h_r and sunset h_s can be computed by:

$$[3] \quad \alpha = \sin^{-1} [\sin \phi \sin \alpha + \cos \phi \cos \alpha \cos h]$$

where ϕ is the latitude; α is the declination of the sun, given by:

$$[4] \quad \alpha = 23.45 \sin (360\ n/365.25)$$

where n is the number of days after March 21 and h is the hour angle of the sun and is positive in the AM and negative in the PM. The angle of the sun, α is zero at sunrise and sunset and a maximum at noon, h_n.

Therefore, the hour of sunrise can be calculated by solving for h with $\alpha = 0$ in equation [3]. Thus the hour of sunrise is given by

$$[5] \quad h_r = 12 - \{[\cos^{-1}(-\sin\phi \sin\alpha / \cos\phi \cos\alpha)]/15\}$$

The maximum number of daylight hours, h_{sm} for any day is twice the length of time between h_r and h_n, because the solar angle is approximately symmetrical about noon. Thus,

$$[6] \quad h_{sm} = 2(12-h_r)$$

As expected, the maximum hours of sunshine changes with time and latitude because α is a function of the time of year and ϕ defines the latitude.

Given a solar constant, the amount of radiation received at the earth's surface at any instance is a function of the solar angle. Thus, at low angles the amount of radiation received would be less because the length of travel through the atmosphere would be greater and more attenuation would occur. For Edmonton, during the spring period, measurements (figure 4) suggested that the instantaneous solar radiation H_s, under clear sky conditions, can be related to the solar angle by:

$$[7] \quad H_s = 3.7 \sin\alpha \qquad (GJ/hr\text{-}m^2)$$

Combining equations [3] and [7]; integrating between h_r and h_n and doubling results in the total daily clear sky radiation

$$[8] \quad H_d = 2 \int_{h_r}^{h_n} 3.7 [\sin\phi \sin\alpha + \cos\phi \cos\alpha \cos(h\,\pi/12)]dh$$

where the hour of the day has been converted to radians.

Performing the integration produces the total maximum clear sky radiation which can be received on a horizontal surface for any given day.

$$[9] \quad H_d = 7.4 \sin\phi \cos\alpha (12-h_r) + 7.4 \cos\phi \cos\delta (12/\pi) \sin[\pi - (\pi h_r/12)] \quad (GJ/m^2)$$

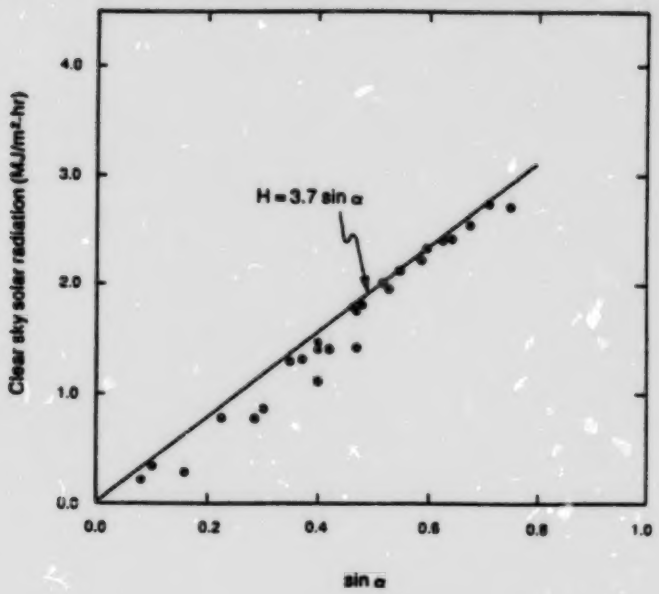


Figure 4 Relationship between the clear sky solar radiation received on a horizontal surface and the solar angle at Edmonton during March and April

To compute daily short wave solar radiation on the basis of hours of sunshine requires correlation between the fraction of daily sunshine and the fraction of maximum potential solar radiation for any given day. Data for Edmonton, for spring conditions, suggests that

$$[10] \quad H/H_d = 0.65\ (h/h_{sm}) + 0.39$$

where H is the measured total daily solar radiation and h is the total recorded hours of bright sunshine. This equation appears to be physically correct. When there is no bright sunshine, the diffuse short wave radiation is about 40% of maximum. When the measured bright sunshine is 100% of maximum, the measured total short wave radiation is 104% of maximum. However, there will be a significant amount of error when applying this equation (figure 5) because the relationship has r^2 values of only 0.75. Nevertheless, it is the only simple tool presently available to estimate the solar radiation over large areas.

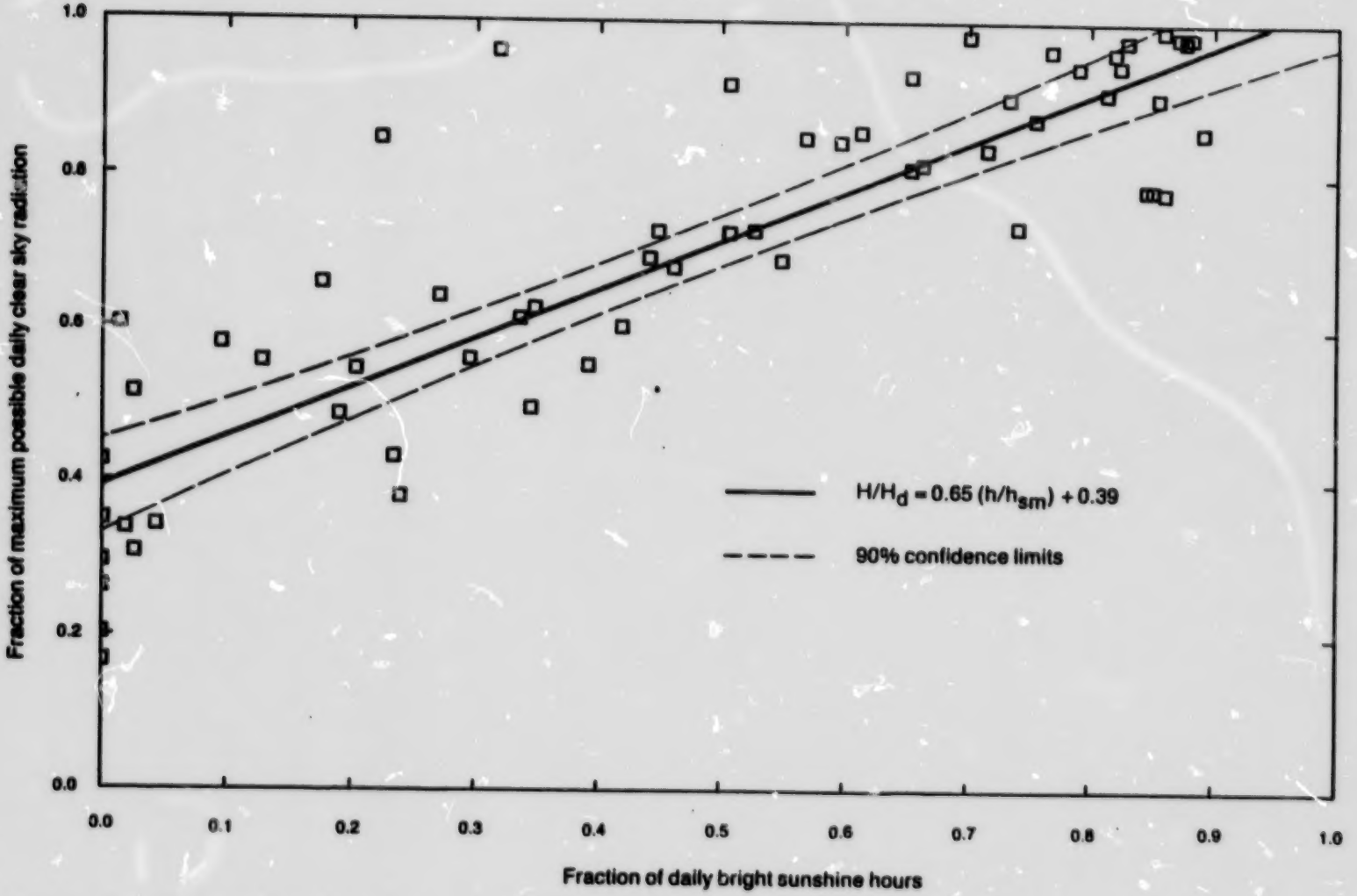


Figure 5 Relationship between the fraction of daily bright sunshine and the fraction of maximum total daily clear sky solar radiation for Edmonton during March and April.

2.3.2 Average Basin Solar Radiation

Figure 3 illustrates the variation of the solar radiation on the Athabasca River basin with time throughout the breakup period. The daily solar radiation fluctuated significantly, from day to day, with the variation averaging about 50% of the mean over a number of days. As expected, there was a significant increase in the solar radiation as the days got longer. Prior to the vernal equinox (March 21), the upper limit of the daily total solar radiation remained relatively constant at about 12 GJ/m^2. However, after that day, there was a significant increasing trend, so that by the middle of April, the upper limit of total daily solar radiation was about 22 GJ/m^2.

2.4 River Discharge

Estimates of discharge throughout the study area were made on the basis of occasional WSC measurements but mostly from daily discharge estimates. Recognizing that under-ice estimates of flows are rather difficult to make, especially during a transitional period such as breakup, there will be some error associated with this analysis.

The Athabasca River was separated into three reaches: upstream of the Pembina River, between the Pembina River and the Lesser Slave River and between the Lesser Slave River and the Clearwater River. This latter reach is extremely long, however, very limited inflow appears. In fact, the techniques for estimating the discharge cannot account for these inflows. This is apparent when comparing the accumulated runoff (figure 6) at Athabasca to that upstream of the mouth of the Clearwater, which was determined by subtracting the Clearwater River inflows from the measured discharge of downstream of Fort McMurray. After April 2, the data suggests there was more flow at Athabasca than at Fort McMurray. This condition persisted until about April 28. No rational way of explaining this discrepancy over such a long time period is apparent, other than that it can be attributed to errors in estimating the daily flows at both Athabasca and Fort McMurray.

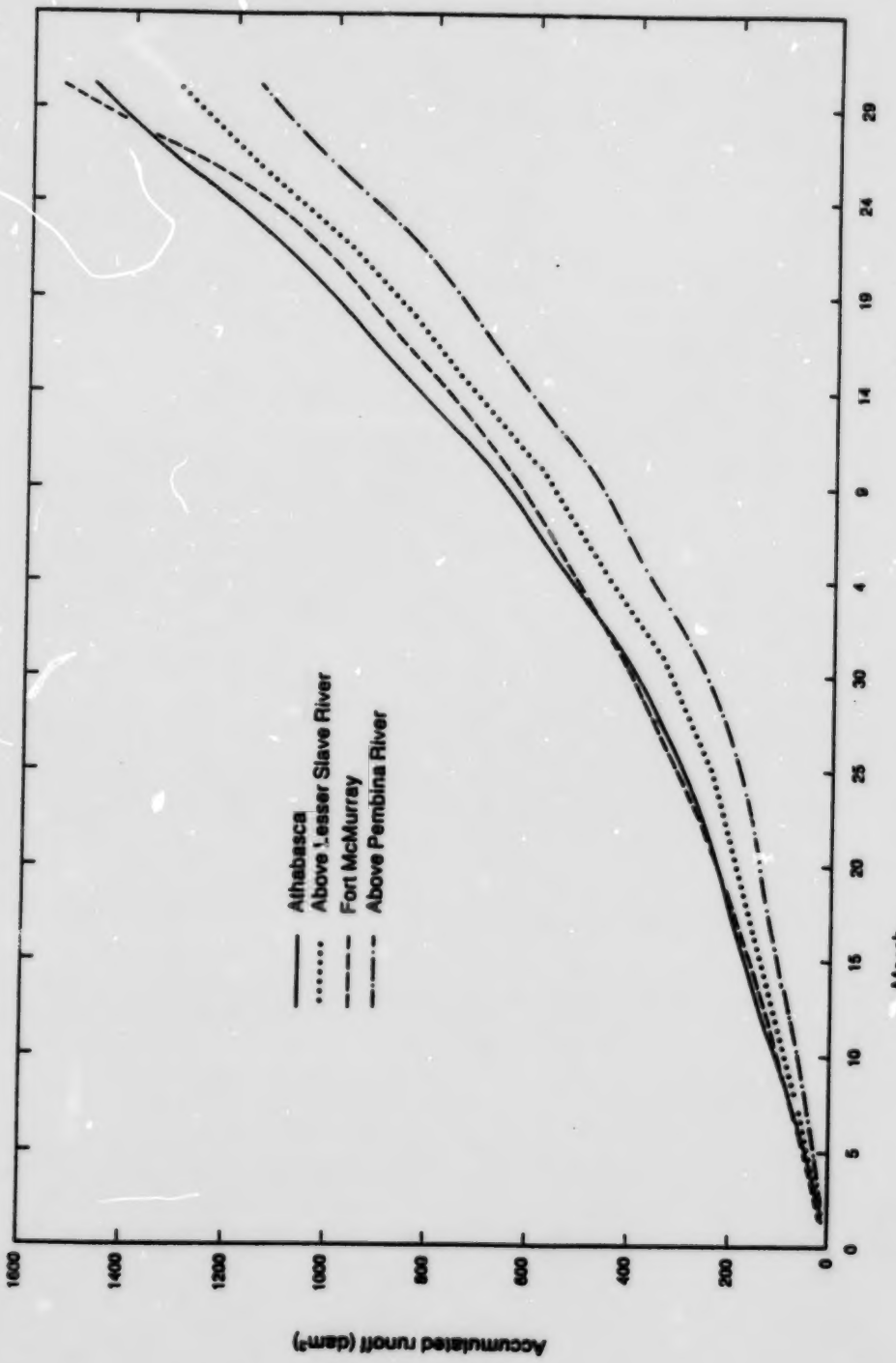


Figure 6 Accumulated runoff on three reaches of the Athabasca River, March and April 1986.

There appears to be three significant periods of response during the breakup and pre-breakup period. Prior to March 25, all three reaches exhibit steady runoff of 79, 107 and 132 m^3/s, respectively. This is most likely base flow. On March 25, there is an acceleration of runoff in all three reaches. The average daily discharge approximately doubles during this period. It is also apparent that runoff from the upper Athabasca River does not increases as fast as that from the Pembina River, which, in turn, does not respond as fast as the basins downstream of and including the Lesser Slave River. This condition persists until about April 3, when it appears that the bulk of the runoff begins. Again, the three reaches seem to behave identically with a rather uniform distribution of runoff through the basin. This is typical of a low snowmelt year, when only moderate runoff is generated. The results are summarized in table 1.

Table 1 Summary of average discharges along the Athabasca River, March and April 1986

Reach	Discharge (m^3/s) Prior to March 25	March 25 - April 2	April 2 - April 19
Upstream of the Pembina River	79	183	302
Pembina River - Lesser Slave River	107	207	336
Lesser Slave River - Clearwater River	132	249	339

3.0 BREAKUP OBSERVATIONS

The progression of breakup was documented on the Athabasca River between the mouth of the Pembina River and Fort McMurray between April 3, 1986, and April 18, 1986. Five aerial reconnaissance were made, one each on April 3, 8, 12, 16 and 18, with the frequency of the observations increasing as the breakup process accelerated. The purpose of the observations were two-fold. First, it was important to document the impact of tributary inflow on the Athabasca River ice, to note areas of open water and the formation of ice jams, to correlate the running ice with the WSC stage records at various gauges in the system, and to produce a description of the overall breakup processes. The second purpose, more important than the first, was to carefully document the percentages of open water and/or overflow in the reach of interest with a view to calibrating and evaluating the model which will be described later. The intent here was to somehow quantify the processes rather than simply describing the breakup in a qualitative manner.

The sequence of melt and runoff can be described and generalized to some extent using Water Survey of Canada (WSC) observations of stage and estimates of discharge on the tributaries of the Pembina River. Both the Upper Paddle River at Anselmo (above the dam) and the Little Paddle River at Mayerthorpe showed a significant increase in discharge in the period of March 15 to March 20, with peak flows evident on March 30. A similar pattern was observed on the Paddle River below the dam with significant stage increases occurring at Rochfort Bridge at Highway 764 and at Barrhead on March 30. This suggests that snowmelt was occurring prior to March 5. Unfortunately, the snow pillow records failed during this period (figure 2) and do not indicate a reduction in the snowpack.

Following the onset of melt on March 26 (figure 2), flows at all the above stations appeared to be decreasing, somewhat contrary to what one might expect given the significant melt that was being observed. The ice effects, as reported by WSC, seemed to end in the first week of April at all the above stations, except for the Paddle River at Anselmo where ice effects continued until April 17. This may be due to the relatively low

runoff at Anselmo and its inability to generate sufficient stage increase to move the ice out.

Figure 7 summarizes the available gauge height information for the Little Paddle River at Mayerthorpe and the Paddle River at Barrhead. The data at Mayerthorpe does not include the critical period, but at Barrhead it appears that the stage was relatively constant at between 1.0 and 1.1 m prior to March 20, at which time it increased to about 2.4 m without breaking up or adversely affecting the ice cover. An additional stage increase to 2.8 m occurred on March 29, followed by a dramatic decrease to 1.3 m over a five day period. This suggests that breakup probably occurred or was initiated on March 29.

On the Pembina River, a significant discharge increase at Jarvie was first apparent March 21, probably from local inflow caused by the same phenomenon which produced a similar response on the Little Paddle-Paddle River system. The peak discharge of this event occurred on April 10, some ten to fifteen days after peaks were recorded on the Little Paddle and Paddle Rivers. Following this, the flow decreased slightly until April 20, when it began to increase again. Figure 7 illustrates the stage variations on the Pembina River. After fluctuating significantly in early March, the stage peaked at 1.9 m on March 9, gradually declining to 1.4 m on March 18 and, subsequently, increasing to another peak of 3.0 m on March 31. This was followed by a steady decrease in stage to 21 m on April 7, after which there was a sudden increase in stage to 3.45 m on April 8, followed immediately by a stage decrease of 1.75 m over a period of 24 hours. This suggests the passage of a surge of ice and water probably destroyed the ice cover. WSC suggests that all ice effects had been removed by April 10, some two days after the surge had passed the gauge.

During this time, the Athabasca River at Athabasca (figure 7) was showing a very limited response. The stage was relatively constant at 1.30 m until March 16, when a gradual stage increase began to appear. This continued until April 15 when the stage reached 2.3 m. A relatively sudden reduction of 0.6 m, over a period of 48 hours, followed by very

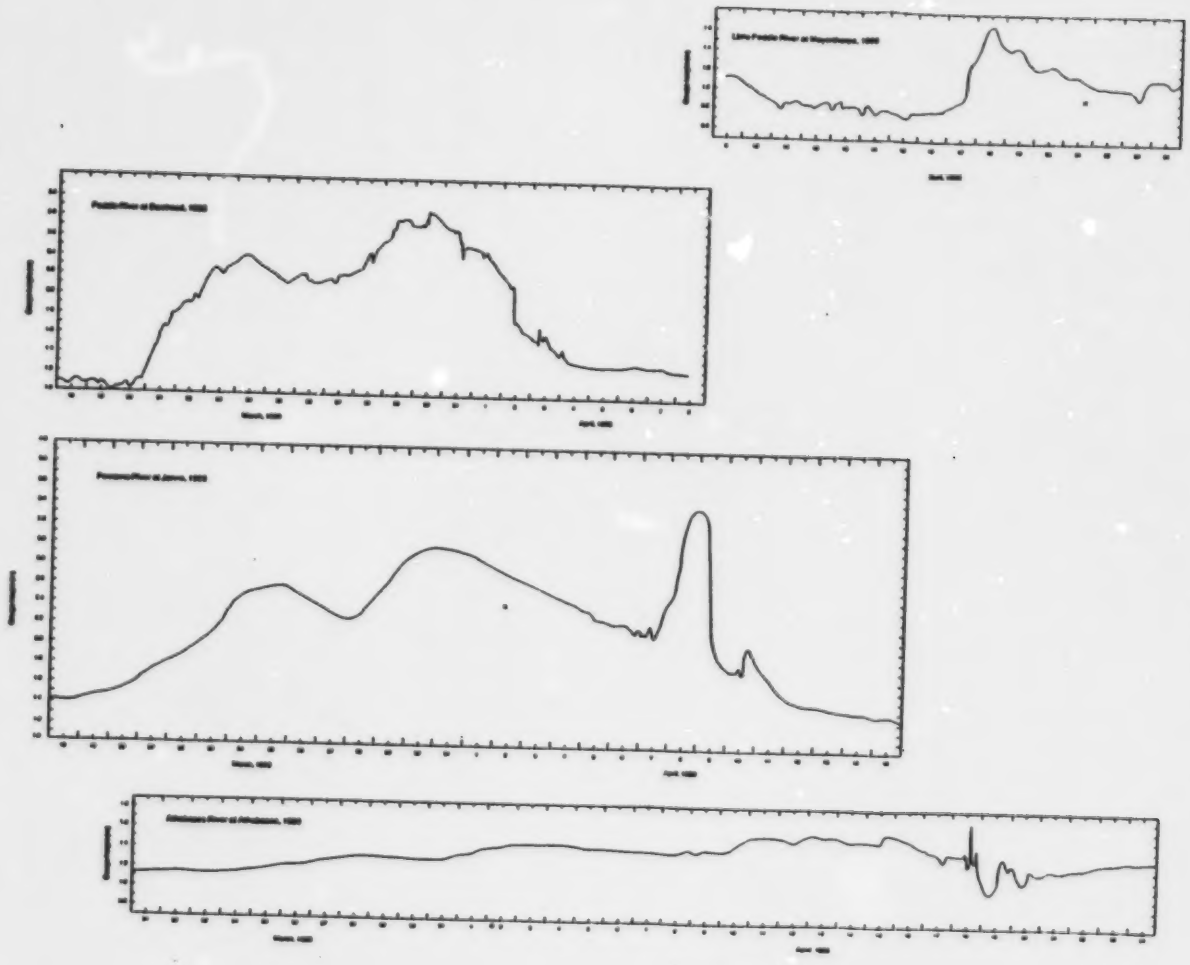


Figure 7 Water levels at selected WSC gauging stations on the Little Paddle, Paddle, Pembina and Athabasca Rivers, March and April 1986

erratic stage fluctuations suggests breakup occurred on April 18. On April 19, a rather symmetrically shaped stage increase was recorded, suggesting that an ice run had occurred and from the smooth stage fluctuations following, it is apparent that the river was cleared of ice. WSC suggests that all effects of ice were removed on April 21.

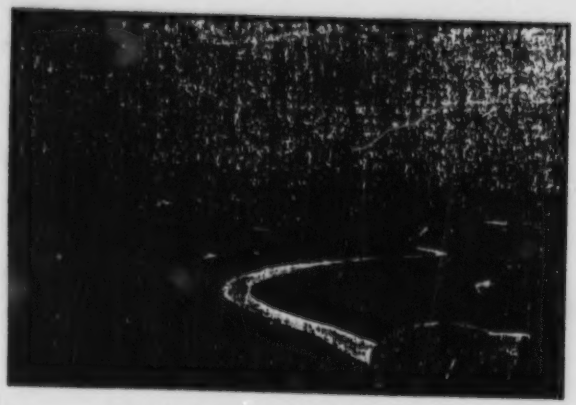
3.1 Ice Characteristics

3.1.1 Conditions on April 3

This reconnaissance extended from Barrhead on the Paddle River, downstream to the Pembina River, which was then followed to its mouth on the Athabasca River. The Athabasca River was surveyed from there downstream to Fort McMurray.

The Paddle River, downstream of Barrhead, was generally open with only short accumulations of broken ice near its mouth. The ice cover on the Pembina River was virtually intact, but it appears that significant border melt and/or an increase on stage had lifted the cover clear of the banks. Some open water was evident downstream of Rossington, but only in the vicinity of the upstream ends of many of the cutoffs. This pattern extended downstream to Fawcett, at which point open water, punctuated by short lengths of solid and broken ice covers, was observed. This instability did not impact on the stage at Jarvie. A significant jam was observed in the reach near Kilsyth. The ice cover at the mouth was intact with the Pembina River having no effect on the Athabasca River.

The Athabasca River ice was solid, except for a short open lead along the right bank, downstream of the mouth of the Pembina River. There was no apparent ice movement, and between the Pembina River and Rourke Creek, only the odd bit of overflow could be observed in the center of the channel. At Rourke Creek (figure 8), there was a small amount of open water on the inside part of the bend, but in general, the ice cover was very competent. Downstream of Rourke Creek, additional overflow was observed near its mouth and downstream of the mouth of Hondo Creek. Overflow was also evident for about 2 km, beginning from a point



April 3, 1986; looking downstream

April 8, 1986; looking downstream

April 12, 1986; looking downstream

April 16, 1986; looking upstream

April 18, 1986; looking downstream

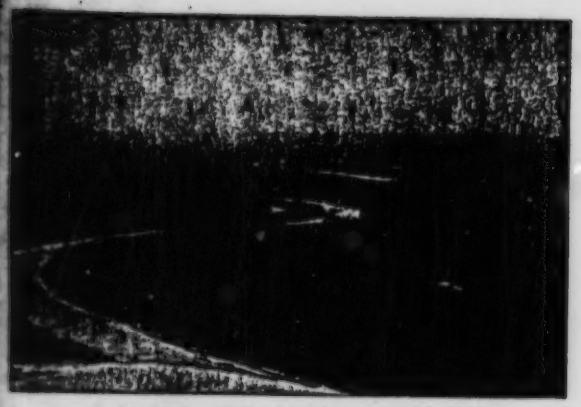
Figure 8 Deterioration of the ice cover and progression towards breakup on the Athabasca River at Rourke Creek

1 km downstream of Highway 2.

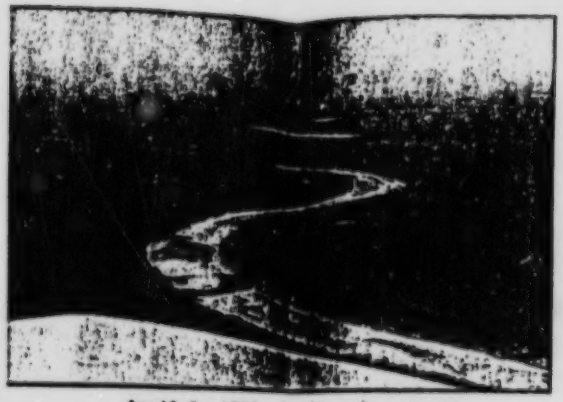
The ice at the mouth of the Lesser Slave Lake was intact with no overflow or open leads evident on the Athabasca River. Some border flow was observed at the bridge crossing at Smith. This pattern dominated as far downstream as the mouth of Tomato Creek, at which point the ice cover appeared to be a little more deteriorated. The photographs in figure 9 are typical of the ice cover in this reach. Open leads became more predominant downstream of this point, and it was apparent that outflow from most of the small tributaries produced border melt on the Athabasca River at least for a short distance downstream of the tributaries. The situation at Athabasca is illustrated in figure 10.

Downstream of Athabasca, the ice cover characteristics did not change significantly until near the mouth of the La Petite Riviere Jaillante, where the cover became more deteriorated with significant mid-channel leads and overflow being more evident. This overflow was evident in the vicinity of the large island WNW of Corrigall Lake (figure 11), but downstream of there, the cover remained intact. At the mouth of Duncan Creek, overflow had created some open water, but the cover regained its solid appearance in the reach west of MacMillan Lake where only discontinuous border flow was evident. Iron Point (figure 12) showed signs of both overflow and open water for a distance of about 8 km downstream.

North of the 20th base line, the ice cover was intact and solid. However, considerable deterioration, probably from overflow, had occurred within a 3 km long reach located some 10 km downstream of that point. A similar type of deterioration was observed at Pelican Point, Pelican Rapids and Stony Rapids, although these open leads were relatively narrow and discontinuous. Overflow was also evident downstream of Stony Rapids, but below this stretch of overflow, the ice appeared intact as far downstream the hairpin bend upstream of Rapides du Joli Fou, where some deterioration and opening had occurred. Upstream of these rapids, narrow mid-channel open leads were evident and downstream of the rapids, only overflow was observed. The rapids were intact, however. From there,

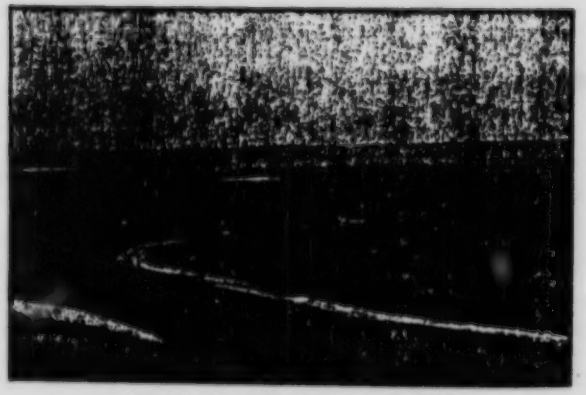


April 3, 1986; looking downstream



April 8, 1986; looking downstream

April 12, 1986; looking downstream



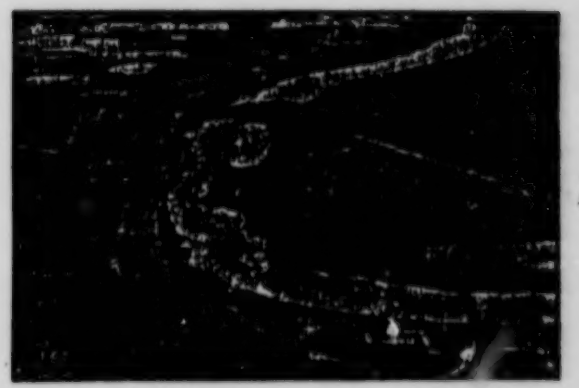
April 16, 1986; looking upstream

April 18, 1986; looking downstream

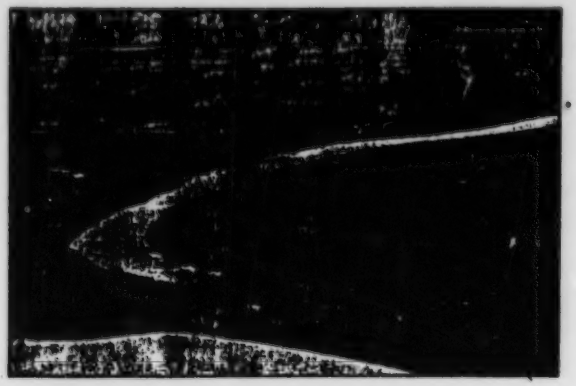
Figure 9 Deterioration of the ice cover and progression towards breakup on the Athabasca River southwest of Calling Lake

April 3, 1986: looking downstream

April 8, 1986: looking downstream



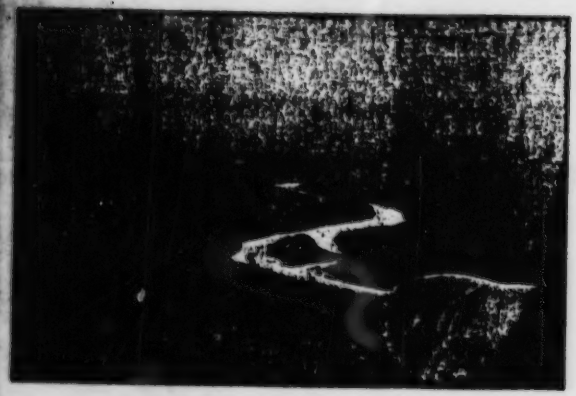
April 12, 1986: looking downstream



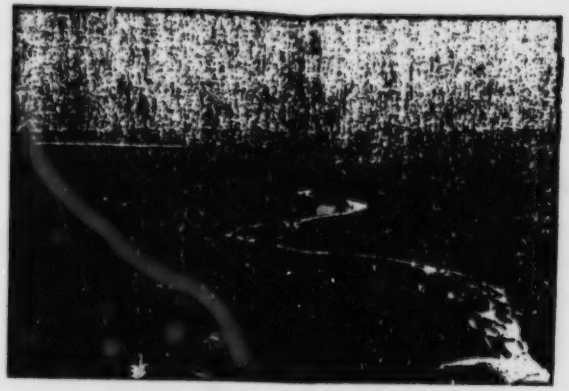
April 16, 1986: looking upstream

April 18, 1986: looking downstream

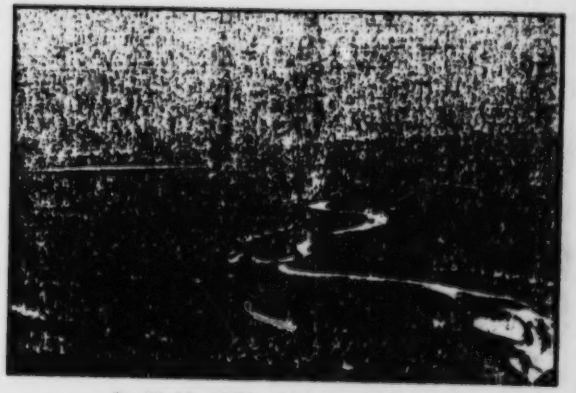
Figure 10 Deterioration of the ice cover and progression towards breakup on the Athabasca River at Athabasca



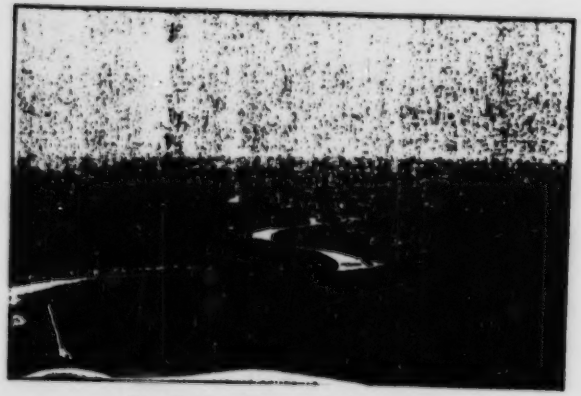
April 3, 1986; looking downstream



April 8, 1986; looking downstream



April 12, 1986; looking downstream



April 16, 1986; looking upstream

April 18, 1986; looking downstream

Figure 11 Deterioration of the ice cover and progression towards breakup on the Athabasca River west of Corrigall Lake

April 3, 1986; looking downstream

April 8, 1986; looking downstream

April 12, 1986; looking downstream

April 16, 1986; looking upstream

April 18, 1986; looking downstream

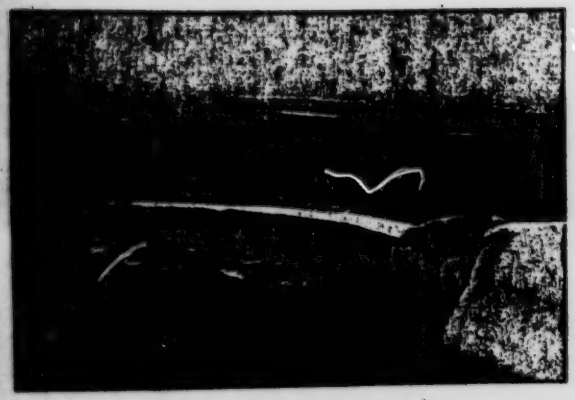
Figure 12 Deterioration of the ice cover and progression towards breakup on the Athabasca River at Iron Point

downstream to Grand Rapids, the cover was intact. The intact ice cover at the mouth of the House River is shown in figure 13. At the rapids, the cover was somewhat deteriorated with the rapids just beginning to open up. Discontinuous open leads had formed for a distance of about 3 km downstream of the rapids.

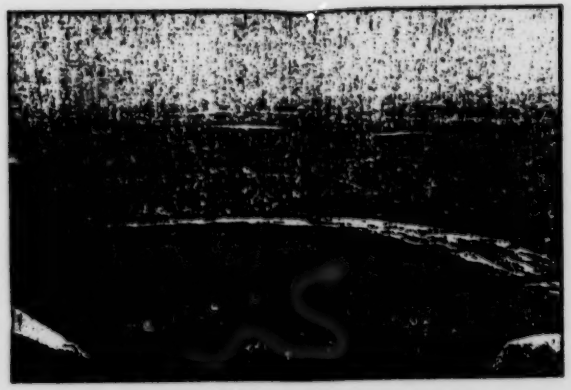
Downstream of Point La Biche and past Brule Point (figure 14), the cover was solid, however, Brule Rapids were just beginning to open up, and considerable overflow was apparent. Downstream, the ice cover appeared solid, except for overflow at the Gooseneck and upstream of the Alger River and open leads through Long Rapids. Crooked Rapids showed some overflow (figure 15) and the ice cover at Rock Rapids was deteriorating. Overflow was evident through and downstream of Cascade Rapids for a distance of about 8 km. Upstream of Mountain Rapids, the cover was intact, with the rapids just beginning to open up and exhibiting overflow. A significant amount of overflow was also observed between the island downstream of Mountain Rapids and Moberly Rapids. Downstream of Fort McMurray, the entire left hand side of the channel (within the islands) seemed to be either open or subjected to overflow.

3.1.2 Conditions on April 8

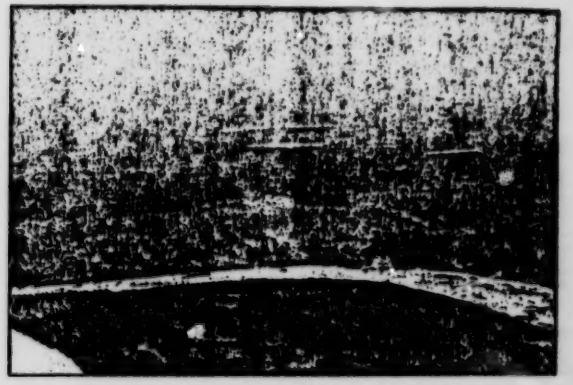
The extent of this reconnaissance was similar to that on April 3. By this date, over a period of five days, the Paddle River was entirely ice free and upstream of the Paddle River, the Pembina River had opened up completely. There was some evidence that a small run, less than 1 m in height, had occurred. A short stretch of running ice at Manola and a much larger run extending from 3 km downstream of Freedom to 3 km downstream of Rossington were observed within the long reach of open water leading up to the head of a jam at the upstream end of the cutoff located about 2 km downstream of the Pibrock Bridge. This ice jam extended downstream to within 2 km of the Dapp Bridge where the head of another accumulation was observed. This jam extended as far downstream as the mouth of Shoal Creek, downstream of which the channel was alternately open and ice covered due to the presence of short accumulations. This condition prevailed as far downstream as Jarvie, at which point, the ice



April 3, 1986; looking downstream



April 8, 1986; looking downstream

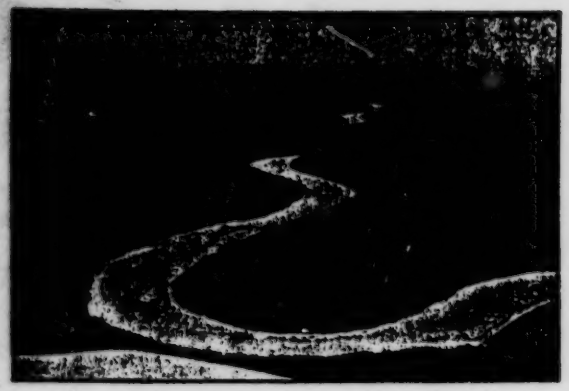


April 12, 1986; looking downstream

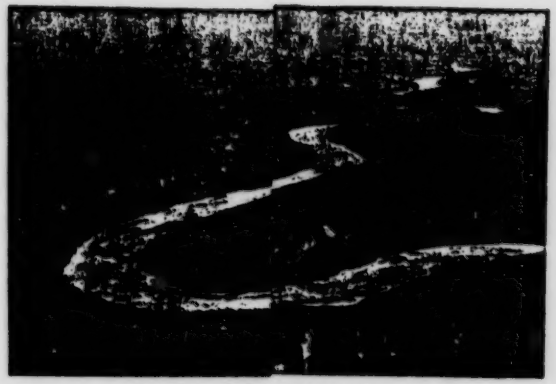
April 16, 1986; looking upstream

April 18, 1986; looking downstream

Figure 13 Deterioration of the ice cover and progression towards breakup on the Athebasca River at the mouth of the House River



April 3, 1986; looking downstream

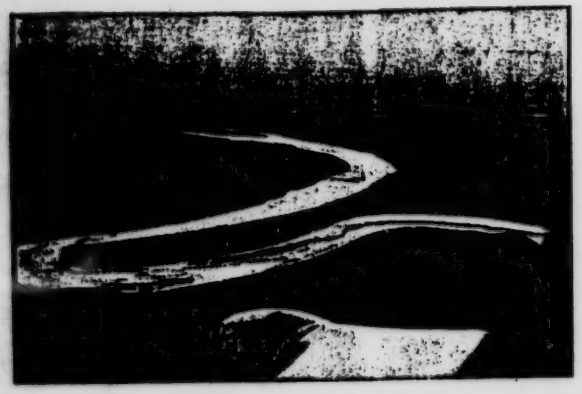


April 8, 1986 looking downstream

April 12, 1986; looking downstream

April 16, 1986; looking upstream

Figure 14 Deterioration of the ice cover and progression towards breakup on the Athabasca River at Brule Point



April 3, 1986; looking downstream

April 8, 1986; looking downstream

April 12, 1986; looking downstream

April 16, 1986; looking upstream

Figure 15 Deterioration of the ice cover and progression towards breakup on the Athabasca River at Crooked Rapids

cover became solid and generally intact, but rotting significantly. Downstream of Jarvie, the cover characteristics had not changed significantly from the observations on April 3. Sporadic open water and short lengths of broken ice were still observed at Kilsyth, and the ice cover at the mouth of the Pembina River was still intact.

The record of stage at Jarvie (figure 7) suggests that at the time of this observation, discharge was significantly increasing. Although direct observations cannot confirm this notion, in all probability the ice cover at Jarvie was destroyed late on this day or on April 9. Additional increases in stage on April 10 probably resulted from the passage of the ice run observed on this day upstream of Rossington.

The Athabasca River had deteriorated considerably since the last observations. Numerous small open areas had developed, and many of the previously covered gravel bars were now exposed. The open lead downstream of the mouth of the Pembina River almost doubled in length, whilst upstream of Chisholm, the cover is intact., At Chisholm transverse cracking had occurred, leading to considerably more overflow and ultimately the formation of open leads. Downstream of Chisholm, the cover appeared very disintegrated and on the verge of becoming unstable, with open areas and overflow becoming the dominant feature.

At Rourke Creek, slightly more deterioration and more open water was evident (figure 8), but no significant changes were apparent between Rourke Creek and the Highway 2 bridge at Hondo. At the bridge, an open lead, extending almost to the island downstream, had developed in the center of the channel. The Lesser Slave River was still ice covered at the mouth and the Athabasca River ice cover appeared deteriorated, but still intact upstream and at Smith. Downstream of Smith, the ice cover was also intact, but with very strong and well defined border melt.

Upstream of Tomato Creek, there was a dramatic change in the ice cover (figure 9). The previously observed overflow has developed into open water. Numerous transverse cracking had occurred, giving rise to more overflow, resulting ultimately in open water. However, there was no

obvious downstream movement of the ice cover. The previously observed overflow located downstream of Tomato Creek had disappeared, leaving instead disjointed areas of open water in the center of the channel. Further downstream, transverse cracking was apparently generating some overflow.

Downstream of a line running east from Lawrence Lake, the cover was once again solid for as far downstream as the mouth of the creek draining Mud Lake. At this point, additional transverse cracking and open water was observed in a cover that was relatively solid, but more deteriorated than on April 3. Upstream and in the vicinity of Athabasca (figure 10), very little change in the ice cover was observed although it appeared somewhat more deteriorated, as one would expect.

Downstream of Athabasca, a large open lead developed from a point 1 km downstream of the bridge and transverse cracking was observed downstream around the bend where overflow was documented previously. Downstream of the mouth of Sawdy Creek, the ice cover was solid but continuing to deteriorate. This characteristic extended downstream to a point 6 km downstream of Pleasant View, at which point numerous open leads became evident for a distance of about 5 km.

Upstream of the La Biche River, the cover appeared solid except for local deterioration around the confluence. Downstream, the cover was intact. This extended to the mouth of the Calling River where a 2 km long open lead had developed. Downstream of the Calling River, the cover was solid except for transverse cracking at the mouth of La Petite Riviere Jaillante, downstream of which the open leads previously observed had become more continuous. These open leads extended downstream to the island NW of Corrigall Lake (figure 11), at which point the cover appeared to become more competent for at least as far downstream as Iron Point.

At Iron Point (figure 12), considerably more open water was observed and a deteriorating cover existed downstream of there for about 3 km. The long lengths of overflow previously observed were not evident, although the ice cover was left mottled and much darker in colour.

Approximately 10 km downstream of Iron Point, a 3 km long open lead had developed and downstream of there, open leads and transverse cracking prevailed to a point 5 km north of the 20th baseline. Below this point, 2 km of ice had moved downstream into a 2 km long open lead. Below this, the ice cover was solid, but with some transverse cracking leading up to a 4 km long open lead upstream of Parallel Creek.

At Pelican Portage, the ice cover was intact, but deteriorating, and the discontinuous open leads along the thalweg of Pelican Rapids had not changed since the previous observation, except that Stony Rapids exhibited some areas of open water. These leads, which were substantially a product of border melt, extended about 6 km downstream of the rapids. From there, downstream to the bend located 6 km upstream of Rapides du Joli Foi, the cover was generally solid, but at this point, more open leads were evident. At the rapids, there was a 3 km long open lead in existence upstream, but a relatively solid looking cover downstream. Between there and the House River, the cover had deteriorated considerably since the last observation, and a wide transverse rack had formed 2 km upstream of the House River confluence (figure 13).

Downstream of the House River, the cover was solid looking and intact. Grand Rapids had opened up considerably with open water extending from a point 1 km upstream of the rapids to 6 km downstream. The cover at Point La Biche was solid (both upstream and downstream), but a transverse crack was evident at a location 4 km downstream. This produced some open water which had previously been noted as overflow. Between there and Brule Rapids, the cover was solid except for a transverse crack located 3 km upstream of Brule Point and open leads with a deteriorated cover just downstream of there (figure 14).

At Brule Rapids, the leads previously observed were still evident, but not significantly greater in size. Between Brule Rapids and the Gooseneck, the cover appeared solid. At the Gooseneck, a section of open water had developed, but downstream, the cover was still solid. This solid cover extended downstream to a point 5 km upstream of the Alger River, where a lead had developed along the right bank. This lead

extended downstream past the confluence and developed into a 2 km long area of large open leads and deteriorated cover. Sporadic areas of open water were observed within Boiler Rapids, upstream of which there was considerable overflow for a distance of about 3 km. Overflow and open water was evident over 10% to 20% of the channel width in the reach between Middle Rapids and Long Rapids. At the latter rapids, a large open hole was observed on the inside of the bend along the right bank, and open water was observed to stretch continuously for 5 km downstream. From there, downstream to Crooked Rapids, considerable open water was again evident upstream of the sharp bend. Downstream of there, open water was evident at Rock Rapids (figure 15), with some open spots between Little Cascade and Cascade Rapids. Mountain Rapids had deteriorated somewhat more since the last observation, and a narrow lead had formed for about two channel width upstream and downstream of the islands below Mountain Rapids. Below this lead, the cover was solid past Moberly Rapids and through MacEwan Bridge. The overflow between the islands along the left bank near the mouth of the Clearwater River had developed into open water which extended downstream past the WSC gauging station.

3.1.3 Conditions on April 12

Observations on the Pembina River revealed that the river was open, bank to bank upstream of Manola, downstream past Dapp to just south of Fawcett, where a series of short accumulations, punctuated by open water, extended as far downstream of a line due west of Fawcett. It appeared that the majority, if not all of the ice observed on April 8, had melted. From a point due west of Fawcett, the channel was completely open, with the open water leading up to a 4 km long jam whose toe was pushed up against the Athabasca River ice.

On the Athabasca River, the ice cover was severely fractured, but still intact. The holes observed on April 8 had become much larger and more numerous. Numerous transverse cracks, evident for a distance of 10 km downstream of the Pembina River were leading up to open water (bank to bank) which extended downstream to line due west of Bissett Lake. This

ice must have melted in place, because downstream of Bissett Lake, the ice cover was generally intact. It is probable that this open water was generated as a result of previously observed transverse cracking at Chisholm.

The intact ice cover between Bissett Lake and Highway 2 at Hondo was generally not severely fractured, but lengths of open water, in the order of 2-3 channel widths in length, were observed at the locations where overflow and small open holes were previously noted (figure 8). The mid-channel lead in the vicinity of the Highway 2 bridge had increased substantially in length and width, but downstream to Smith, the cover looked reasonably competent.

Near the mouth of the Lesser Slave River, the cover was severely fractured, and a large open section had developed at the mouth. Downstream of the open Lesser Slave River, the channel was alternately ice covered and open for a distance of 12 km to the south of Moose Portage. Downstream of here, the cover was intact and generally solid with no open areas evident, except near the occasional island. This solid cover continued until the river turns southward, west of Calling Lake (figure 9), at which point reaches at open water of about 10 channel widths in length became apparent in the vicinity of Tomato Creek. Downstream of Tomato Creek to Athabasca, the cover was generally intact, with small sections of open water (1-2 channel widths in length) at locations where transverse cracking was previously observed. At Athabasca (figure 10), patches of open water were now evident.

Downstream of Athabasca, there was not much change with the exception that the previously observed open leads had increased in size. These were randomly located and only at a point about 6 km downstream of Pleasant View, did the leads become long narrow and more continuous.

Immediately downstream of the mouth of the La Biche River, a 2 km length of ice force channel was evident, but after that, the ice cover was essentially intact as far downstream as the mouth of La Petite Riviere Jaillante where the transverse crack evident earlier had increas-

ed in size. Open leads which previously were observed upstream and in the vicinity of the islands northwest of Corrigall Lake, had increased in size (figure 11), but downstream of there to Iron Point, the cover appeared competent. The open water observed previously at Iron Point (figure 12) was still evident, as were the open leads downstream. At the 20th base line, continuous open water was encountered from a point 3 km upstream to a location 8 km downstream, below which a 4 km long accumulation was followed by a long narrow mid-channel open lead extending almost to the mouth of Parallel Creek.

From Pelican Portage to the mouth of the House River, the leads produced from border melt and linked by mid-channel leads were continuously growing in extent, although the ice cover appeared to be intact even though its colour suggested a significant deterioration. The transverse crack upstream of the House River had developed into a significant open area. At the mouth of the House River, the cover was intact, but steadily deteriorating (figure 13).

Downstream of the House River, the ice cover continued to deteriorate with large open areas becoming evident. The open leads at Grand Rapids had increased in size and length. Downstream of these leads, the cover appeared to have deteriorated significantly since the previous observations. The transverse cracks, previously noted downstream of Point La Biche, had degenerated into fairly large openings (1-2 channel width in length). Random areas of open water were apparent in the reach between Pointe La Biche and the Gooseneck. At Brule Point (figure 14), the open leads and overflow finally developed into a patch of open water. However, on the whole, no significant increase in the amount of open water was evident since the last observations were made. The cover did seem to be deteriorating, however.

The small patch of open water at the Gooseneck had developed into a long bank-to-bank open stretch extending downstream to the Algar River, to the head of a 4 km long accumulation with its toe located at Boiler Rapids. Downstream of here, the cover exhibited a rather long, almost continuous narrow open lead from Boiler Rapids, through Middle Rapids to

a point just downstream of Long Rapids. From there to Crooked Rapids the cover was intact, at which point considerably more open water was apparent since the last observation (figure 15). This open water extended downstream to Rock Rapids, below which the ice cover was intact except for the occasional open area and numerous, small disjointed open leads which were more concentrated in the rapid sections. There was, however, no significant change in the cover characteristics from that observed on April 8.

3.1.4 Conditions on April 16

By this date, the Pembina River was completely clear of ice with only a 1 km long accumulation of ice still evident at its mouth. Upstream of the Pembina River, the Athabasca River was still ice bound, as was a reach of river about 10 km long downstream. At that point, due west of Tieland, the river was open (bank to bank) to a location about 5 km downstream of Chisholm, where a very deteriorated ice cover was encountered. This cover persisted as far downstream as Rourke Creek, at which point alternate segments of open water and an intact ice cover were observed (figure 8). This condition prevailed as far downstream as about 3 km below the Highway 2 bridge at Hondo where a solid, but deteriorating ice cover was again observed. At Smith, this evolved into open water, which extended continuously to just south of Moose Portage. In this reach, approximately one half the channel (mostly along the left bank) was open. At Moose Portage, more ice and open water (in alternating blocks) were observed, and this condition (figure 9) extended downstream to Tomato Creek, at which point the ice cover became more continuous. This continuous, more competent ice extended downstream to east of Howie Lake, when again the alternating blocks of competent ice and open water became evident. This condition persisted downstream (except for a 10 km long intact reach located downstream of Baptiste Creek) as far as the mouth of Sawdy Creek. At Athabasca (figure 10), considerably more water was evident at the upstream end at the town. However, the ice upstream of the town was intact. The stage record (figure 7) did not suggest incipient breakup on this day. About 1 km of ice was lodged against the bridge, downstream at which the channel was completely clear of ice for a

distance of 3 km.

Downstream of Sawdy Creek, the ice cover was generally intact with less blocks of open water and more pronounced long narrow leads. At the mouth of the La Biche River, open water was once again encountered with only an intermittent, highly fractured, transversely cracked, dislodged ice sheet evident along the left bank. This condition extended downstream to the mouth of the Calling River, below which the cover once again became continuous, but with only an occasional patch of open water and the odd narrow open lead. A large ice free stretch, about three channel widths in length, was still evident at the mouth of La Petite Riviere Jailiante, but downstream of there, the channel was essentially ice covered except for a location in the vicinity of the islands northwest of Corrigall Lake (figure 11) and where a few more open leads and patches of open water had formed at the transverse cracks which had developed since the last observation. A relatively intact and competent ice cover was observed between Duncan Creek and Iron Point. At Iron Point (figure 12), the amount of open water had increased somewhat and downstream of there the ice cover appeared to be considerably more deteriorated, with a subsequent increase in the amount of open water. The ice cover, however, was not fractured, nor was it dislodged. At the 20th base line, the open water had extended from a point 3 km upstream to 10 km downstream, a slight increase in length since the previous observation. The accumulation below this open water had decreased in length to about 3 km. Below this, another 4 km long completely open segment of channel was observed, leading up to a 1 km long accumulation also impinging on a solid ice cover which, in turn, extended as far downstream as Pelican Portage. A number of small open leads were apparent within Pelican Rapids, but the cover was intact.

Downstream of Pelican Portage, the cover was again punctuated by numerous open areas and long mid-channel open leads. This was especially evident downstream of Stony Rapids where the cover was quite deteriorated. At the sharp bend located 10 km upstream of Rapides du Joli Fou, the ice cover became almost non-existent with only the odd remanent of intact ice. This condition persisted downstream to the mouth of the House

River, where the open water, from upstream, enveloped the large open area observed previously (figure 13).

Downstream of the House River (figure 13), the ice cover was intact for a distance of about 7 km. This was followed by an entirely open channel which extended past Grand Rapids, to a point 6 km downstream. Below this open water, there was a 4 km long accumulation of broken ice impinging against an intact cover which extended downstream to Point La Biche.

Downstream of Point La Biche, the ice cover was completely removed for a distance of about 10 km. Below this reach, there was a 1 km long accumulation of ice arrested by a solid, but deteriorating ice cover which stretched as far downstream as Brule Point, below which a 2 km stretch of open water was observed. From there, downstream for 6 km, the cover was intact, but very deteriorated, with numerous open holes and mid- channel leads. This was followed by about 3 km of open water upstream of Brule Rapids, below which a very competent cover extended downstream as far as the Gooseneck.

The completely open reach, observed between the Gooseneck and the Alger River, had not changed significantly since April 12, except that it may have lengthened by 1 or 2 km. Downstream of the Alger River, the ice cover appeared to be generally competent, although deteriorating with numerous mid-channel leads and punctuated by a 1 km long open stretch at Boiler Rapids and a 2 km long reach at Middle Rapids. At Crooked Rapids (figure 15), open water had worked its way about 4 km upstream from the rapids, but its downstream extent was limited to Rock Rapids. Downstream of Rock Rapids, very little change was apparent since the observations on April 12. In fact, it seemed as if some of the overflow was somewhat reduced, indicating a reduction in discharge or, perhaps, a freeing up of the ice cover. The only exception was the development of a large lead along the left bank extending from the Syncrude pipeline crossing, downstream to Little Cascade Rapids.

3.1.5 Conditions on April 18

Observations on this date were made along the Athabasca River to Grand Rapids. Unfortunately, low cloud and fog just downstream of Grand Rapids forced the discontinuation of the flight. Therefore, complete coverage of the whole river upstream of Fort McMurray could not be made.

The Pembina River was completely clear of ice, with the jam previously located at the mouth having been melted out or moved into the Athabasca River. On the Athabasca, the ice cover was still intact upstream at the mouth of the Pembina River, but downstream, the cover had deteriorated to a point that continuous open water extended as far as Tieland, downstream of which for 3 km, there was a fractured, rearranged, but still stationary ice cover. Below this, the channel was completely open downstream to a point 6 km below Chisholm. Here, a very fractured and rearranged ice cover, punctuated by open stretches in the order of 1 km long, was evident for as far downstream as the Rourke Creek bend, where the cover remained solid, unfractured and intact (figure 8).

Downstream of the bend, the channel was completely ice free through the bridge section at Highway 2 near Hondo for 5 km, below which a solid cover extended to the downstream end of the large island located upstream of Smith. From that point, the channel was open downstream to a location immediately east of Moose Portage, where a solid cover was encountered. This cover, punctuated by short open stretches (1 to 2 channel width wide) extended downstream for 8 km, below which a 4 km stretch of open water was observed (figure 9), followed by 4 km of ice and 5 km of open water leading up to a 3 km length of ice cover just upstream of Tomato Creek. Downstream of there, the channel was alternately open and ice covered for various distances. Between Tomato Creek and Athabasca, the observed ice cover was segmented into lengths ranging from 1 km to 10 km. Much of the cover was intact and unbroken, but there were some accumulations of large sheets, obviously moved but not acted upon by dramatic or intense surges. At Athabasca, immediately adjacent to the town and upstream of the bridge, the cover was intact but deteriorated significantly (figure 10). This suggests that an ice run had not yet developed.

This observation was made just prior to the start of significant stage fluctuations at Athabasca and probably just prior to the start of the ice being cleared from the channel. These same conditions extended downstream from Athabasca to Pleasant View, at which point open water was encountered for a distance of about 8 km, leading up to a short accumulation of broken ice impinging on a solid but porous and highly deteriorated intact ice cover. This cover extended downstream to the mouth of the La Biche River, at which point the cover appeared to have just broken up and began moving downstream in rather large ice sheets. This short, youthful ice run extended for only about 7 km, moving down into the previously established open water reach upstream of the mouth of the Calling River.

Downstream of Calling River, the ice cover was intact, but continuing to deteriorate, still with only an occasional patch of open water and the odd long narrow lead evident. The opening at the mouth of La Petite Riviere Jaillante was still apparent, and although the ice cover downstream of there was still intact, it had undergone considerable deterioration and numerous transverse cracks had formed since the last observation. In the vicinity of the island located northwest of Corrigall Lake (figure 11), the previously observed open water had increased in size, but no significant changes in the ice conditions downstream were apparent. From Duncan Creek to Iron Point, the ice cover was intact, but the transverse cracks had grown in length to about one-fifth of the river width. Open water was still evident at Iron Point (figure 12). Downstream of there, the ice cover was intact, but becoming increasingly deteriorated with patches of open water becoming more evident.

Open water still was observed to extend from a point 5 km upstream of the 20th base line to a location 10 km downstream, below which there was a short accumulation of broken ice impinging against a 2 km length of solid ice cover, below which open water extended to the mouth of Parallel Creek. Downstream of Parallel Creek, past Pelican Portage, through Pelican Rapids and Stony Rapids, the cover was very deteriorated, but was essentially intact; but with Stony Rapids, and about 7 km downstream of there, being open. From that point, the ice cover was once again intact, but severely deteriorated with numerous narrow leads and blocks of open

water. This pattern extended downstream to the sharp bend 10 km upstream of Rapids du Joli Fou, where open water became more prevalent, punctuated only by short sections (1 to 2 channel widths in length) of intact ice. Open water was observed to extend to a point 5 km below the House River, with short sections of intact ice evident upstream and downstream of Rapides du Joli Fou and at the mouth of Boivin Creek. A 5 km long reach of broken ice cover which was judged to be stationary, but on the verge of mobilization, separated the House River open reach (figure 15) from the Grand Rapids open reach, which now extended to a point only 6 km upstream of Pte. La Biche. The ice cover at Point La Biche was intact and deteriorating. At this point, the observations were discontinued due to the low visibility.

In summary, for this particular day of observation, there was considerable open water, maybe in the order of 40% overall and up to 80% in some reaches. However, the amount of observed broken and jammed ice was almost insignificant. This suggests that most of the processes which lead to the removal of the ice cover were essentially thermal in nature, and most of the ice melted before it could be transported a significant distance.

3.2 Breakup Progression

The observation program to quantify the progression of breakup was limited by what can be observed on the surface of the ice cover. Thus, only parameters such as the fraction of open water, the amount of overflow, the colour or the texture of the ice cover were documented. Of the above, the most important parameter is obviously the fraction of open water. Its estimation allows for the assessment of both the extent of breakup and the rate at which breakup is progressing. The latter characteristic can be inferred because solar radiation is one of the major driving forces of breakup, and the net average absorption of solar radiation increases as the percentage of open water increases because water has a significantly lower albedo than ice.

The fraction of open water and/or overflow was estimated from both

direct observation and by analyzing oblique 35 mm aerial photographs. For any given length of open water, the fraction of the channel width which was deemed to be composed of open water was estimated to the nearest 1/5 of the channel width. For any given reach of interest, the fraction of open water could be accumulated and defined as a fraction of the reach length. The data is shown in table 2.

Table 2 Observed open water on the Athabasca River between the mouth of the Pembina River and Fort McMurray, 1986

Reach (km)	Fraction of Open Water for a Given Date				
	Apr. 3	Apr. 8	Apr. 12	Apr. 16	Apr. 18
Pembina River-Rourke Creek (850-800)	0.01	0.08	0.50	0.52	0.48
Rourke Creek-Tomato Creek (800-750)	0.03	0.04	0.26	0.32	0.50
Tomato Creek-Athabasca (750-700)	0	0.06	0.20	0.22	0.36
Athabasca-Pleasant View (700-650)	0.01	0.02	0.14	0.16	0.40
Pleasant View-Corrigal Lake (650-600)	0.08	0.06	0.10	0.14	0.40
Corrigal Lake-20th Base Line (600-550)	0.01	0.02	0.10	0.08	0.18
20th Base Line-Pelican Ptge. (550-500)	0	0.08	0.34	0.24	0.50
Pelican Portage-House River (500-450)	0.01	0.04	0.04	0.24	0.52
House River-Brule Point (450-400)	0.01	0.06	0.12	0.12	0.42
Brule Point-Long Rapids (400-350)	0	0.04	0.20	0.20	0.38
Long Rapids-Fort McMurray (350-400)	0.01	0.08	0.10	0.06	0.12
AVERAGE	0.02	0.05	0.19	0.21	0.39

From this table, it is apparent that the fraction of open water increases in time, as would be expected. One might also expect that the upstream reaches between the mouth of the Pembina River and Athabasca, because they are further south, would show dramatically more open water than the more northerly sections of the river. This is not readily apparent and, in fact, except for the reach immediately downstream of the mouth of the Pembina River, the section of river between the 20th base line and the House River seems to generate significant open water earlier than any of the other reaches. This may be attributed to its numerous rapids and sharp bends which are evident in this reach.

Figure 16 gives a graphical summary of the accumulated fraction of open water from the mouth of the Pembina River to Fort McMurray for the various days when observations were made. Of interest are the slopes of various segments of the lines for various days. On April 3, the overall slope of the line is very small and essentially constant. This indicates that there was very little open water, and that conditions were very similar throughout the entire reach of the Athabasca River. On April 8, the slope increases significantly in the reach between the mouth of the Pembina River and Tomato Creek. This indicates that open water was being generated at a faster rate in this reach than along the rest of the river. By April 16, although there was little change in the upper reach (km 850-km 750), the reach of river between the 20th base line and Pelican Portage became the most active. This activity was maintained to April 18, at which time incipient breakup conditions had been reached with a very dynamic breakup occurring the next day.

The most interesting thing, when looking at figure 16, is the relative consistency of the open water fraction along the entire river. There were no reaches at the river where massive amounts of open water were produced in isolation from the other reaches. What is more apparent is a uniform and consistent increase in open water with time, rather than in space.

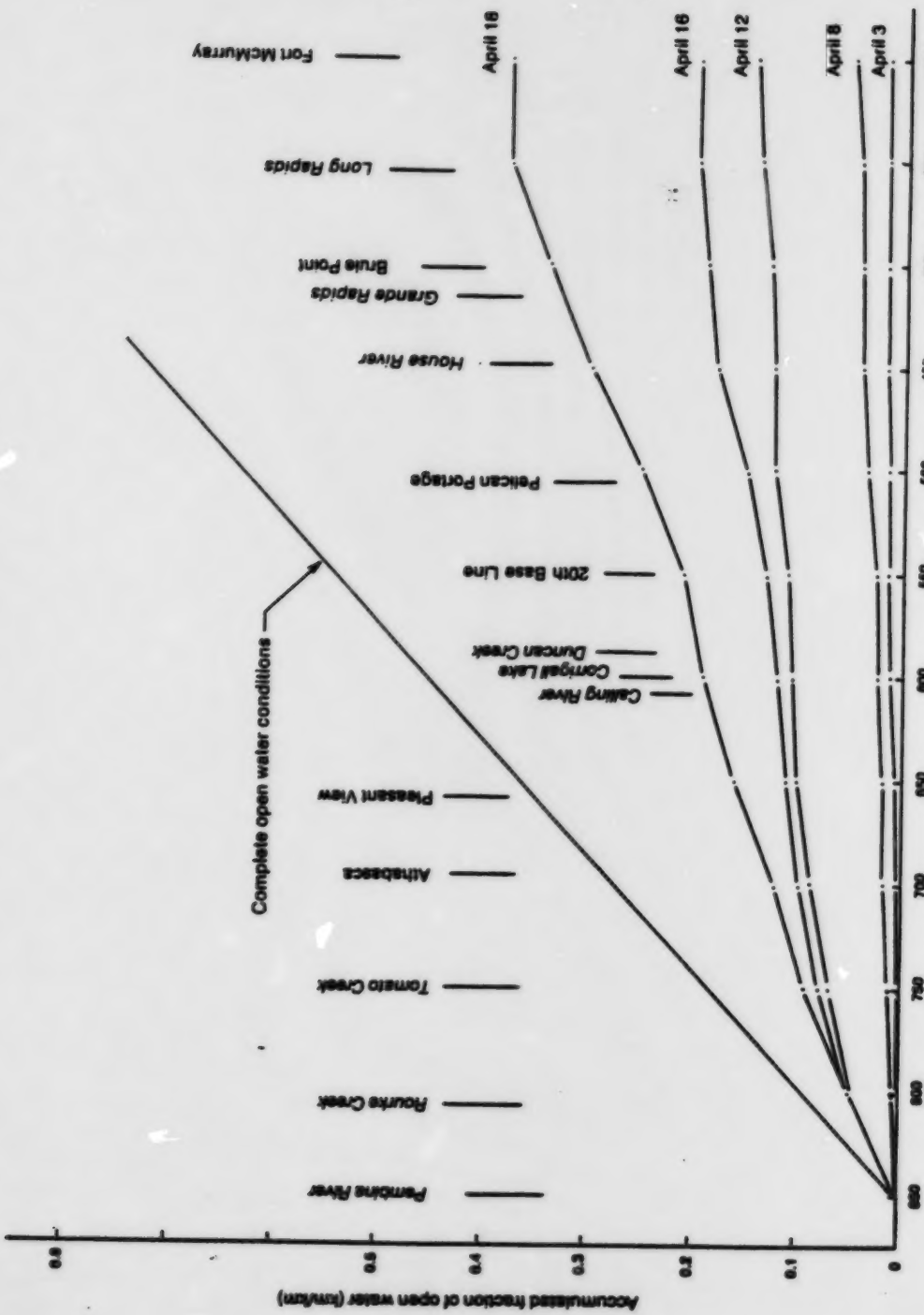


Figure 16 Accumulated fraction of open water on the Athabasca River between the mouth of the Pembina River and Fort McMurray, April 1986

Table 3 relates the progression towards the onset of breakup with some of the climate and discharge information available. As is expected, as the melt progresses, the accumulated solar radiation increases; the degree days of melt increase; the snow on the ground decreases; the accumulated runoff increases and the percentage of open water increases. The difficulty remains to predict the onset of breakup using the readily available data summarized in table 3. The next section attempts to address this problem by briefly summarizing a literature review of existing breakup prediction techniques and describing a model which may be applicable for use on the Athabasca River.

Table 3 Summary of the ice cover and meteorologic data characterizing the breakup on the Athabasca River, 1986

Date	Snow on Ground (mm)[1]	Open Water (%)	Accumulated Solar Radiation (GJ/m²) Beginning March 26	Accumulated Solar Radiation (GJ/m²) Beginning April 3	Degree Days Melt (°C-days)	Accumulated Runoff[4] Beginning March 26 (dam)
March 26	60(44)[2]	>1	0	-	4	0
April 3	13(20)[2]	2	130	0	20	182,000
April 8	4	5	217	87	66	317,000
April 12	0	19	279	149	50 (70)[3]	445,000
April 16	0	21	349	219	46 (80)[3]	589,000
April 18	0	39	380	250	55 (92)[3]	658,000
April 19	0	>50	394	264	61 (100)[3]	693,000

(1) water equivalent at the Twin Lakes snow pillow
(2) brackets indicate basin average
(3) brackets indicate melt discounting sub-zero temperatures April 10 - April 14
(4) measured at Athabasca

4.0 BREAKUP PREDICTION

4.1 Literature Review

The accurate forecasting of the time of breakup on rivers is a significant problem in cold regions around the world. Much of the earliest technology was developed by the Soviets and has been extensively documented by Shulyakovskii (1963). He defines breakup as the first day of "ice drift" or, presumably, the first day of general instability when surges develop and running ice can be observed. In his view, breakup is caused by a combination of the reduction in the ice cover thickness and strength due to melting and the increase in the discharge and river stage which then leads to the physical breaking of the cover. This stage increase causes the ice cover to crack in a variety of locations and eventually to rise free of banks and push downstream. This view is somewhat simplistic because the freeing up of the ice cover does not necessarily create an ice run. Most rivers are meandering and irregular, and the ice gets caught up in bends and on islands and does not move significantly without additional deterioration and/or surges from upstream.

Nevertheless, Shulyakovskii indicates that the amount of heat input, ΣE, which is necessary to cause breakup is

$$[11] \quad \Sigma E = f(t_i, t_s, \phi, v, Y_o, \Delta Y, \Sigma E_b)$$

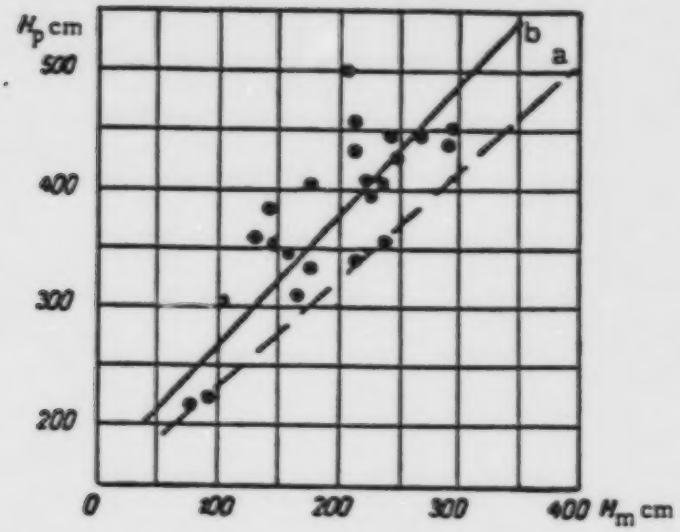
where t_i and t_s are the ice thickness and snow depth, respectively, ϕ is a parameter describing the morphology of the stream, v is the flow velocity at breakup, Y_o is the stage prior to breakup, ΔY is the increase in stage at breakup and ΣE_b is the heat imput at the bottom of the ice cover. Obviously, some simplifications can be made. The ice thickness, t_i, the channel morphology, ϕ, and the pre-breakup stage, Y_o, do not change from year to year. The velocity and snow depth (hence, snowmelt runoff volume) can both be approximated by the stage increase ΔY. The heat input to the bottom of the ice cover is related to the runoff, hence, is also a function of ΔY. This equation thus can be rewritten as:

[12] $\Sigma E = f(\Delta Y)$

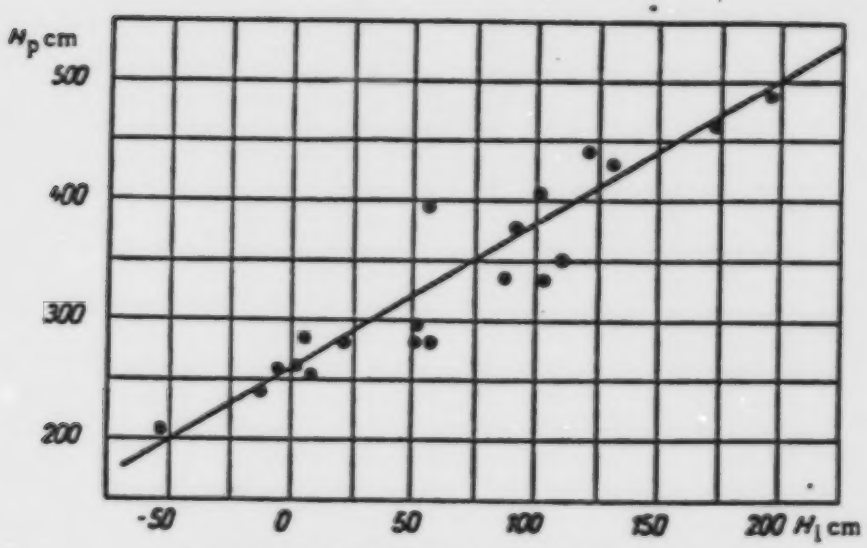
for any given reach of a river. It is important to discuss the implication of ΔY. Under most conditions, the observed range of ΔY for many rivers (as far as it relates to the snowpack runoff) is insufficient to cause the removal of an ice cover. Rather, ΔY is a manifestation of the deterioration of the ice cover, the formation of jams and the resulting surges. This produces large values of ΔY far in excess of that due to the steady inflow from snowmelt.

On the basis of the above rational, Shulyakovskii developed a number of prediction techniques from historical breakup information. The simplest one suggests that breakup occurs when the stage at breakup, Y_p, exceeds the maximum winter stage, Y_m. The ice cover is, thus, freed up and allowed to move downstream. The stage at breakup is also a function of ΣE because the stage increase is a function of the discharge increase which, in turn, is a function of the snowmelt runoff generated upstream. Thus, for any given location and river, by defining Y_p from Y_m (figure 17) and determining ΣE from the $\Sigma E = f(Y_p)$ relationship (figure 18), the onset of breakup can be determined by keeping track of the energy input into the basin. In one particular example, Shulyakovskii plotted the date of breakup against the date on which 30°C-days of melt had occurred in the basin (figure 19). Unfortunately, the scatter is so great ($\pm$5 days over a 20 day period) that the technique is not very useful other than to suggest the earliest possible breakup.

Figure 19 can be modified to account for the northerly advance of the warming trend. In this case, the day of breakup, after the earliest possible breakup in days, was related to the difference in time (days) between the day when 30°C-days of melt was reached in the southern part of a particular basin and the day when 30°C-days of melt was recorded at the location being investigated. To take into account cooling trends, one can also correct the time of breakup by correlating deviations from the earliest date of breakup with the number of days of sub-zero temperatures in the period of time after 15°C days of melt was recorded. Figure 20 illustrates an example.

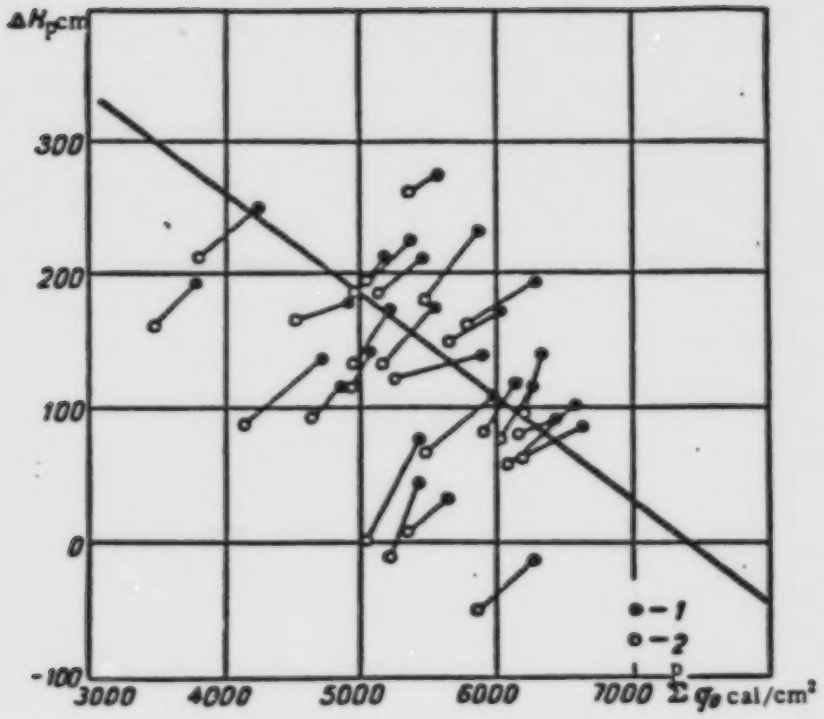


Water stage H_p during ice push on the Lena River at Solyanka as a function of the maximum winter stage H_m.



. Water stage during the first ice push on the Lena River at Krestovskaya, H_p, as a function of the mean water stage during the first 5 days of the stable ice period, H_m.

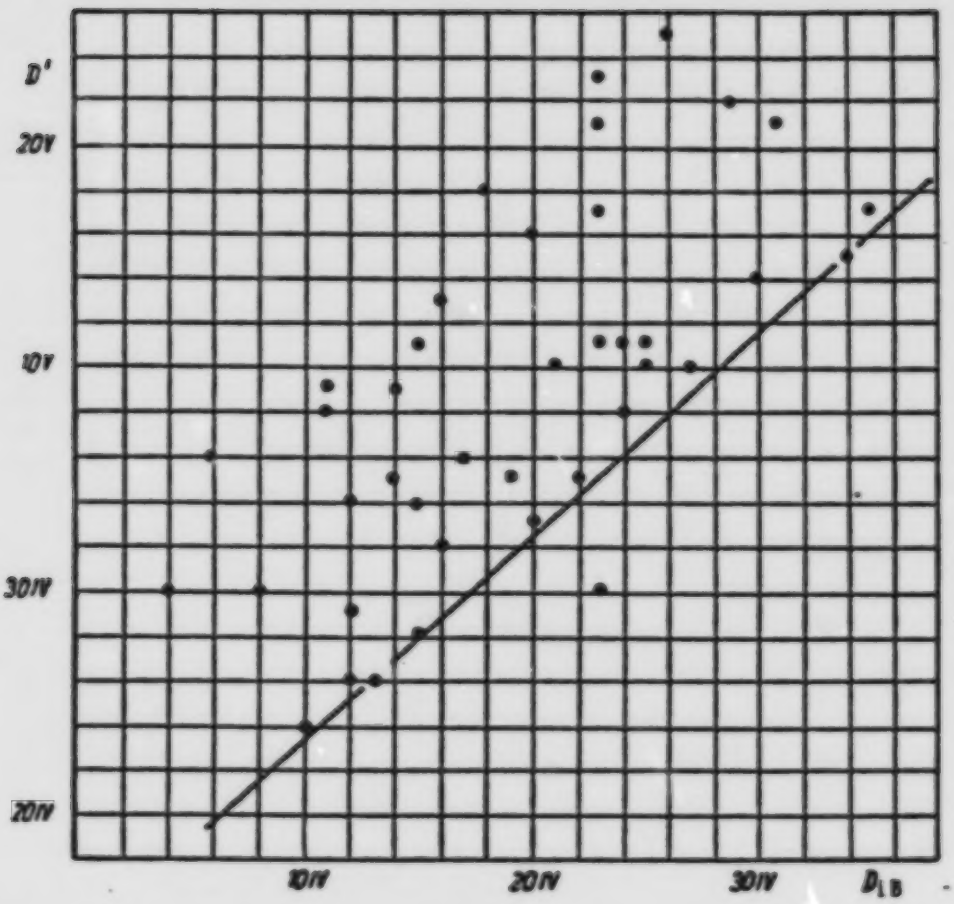
Figure 17 Relationship between the stage increase required for breakup and (a) the maximum winter stage and (b) the mean freeze-up stage (after Shulyakovskii, 1963)



Relationship between the rise ΔH_p of the water stage on the Amur River at Komsomol'sk-on-Amur over the maximum winter stage and the total heat input $\sum q_0^p$ on the day of the first ice push.

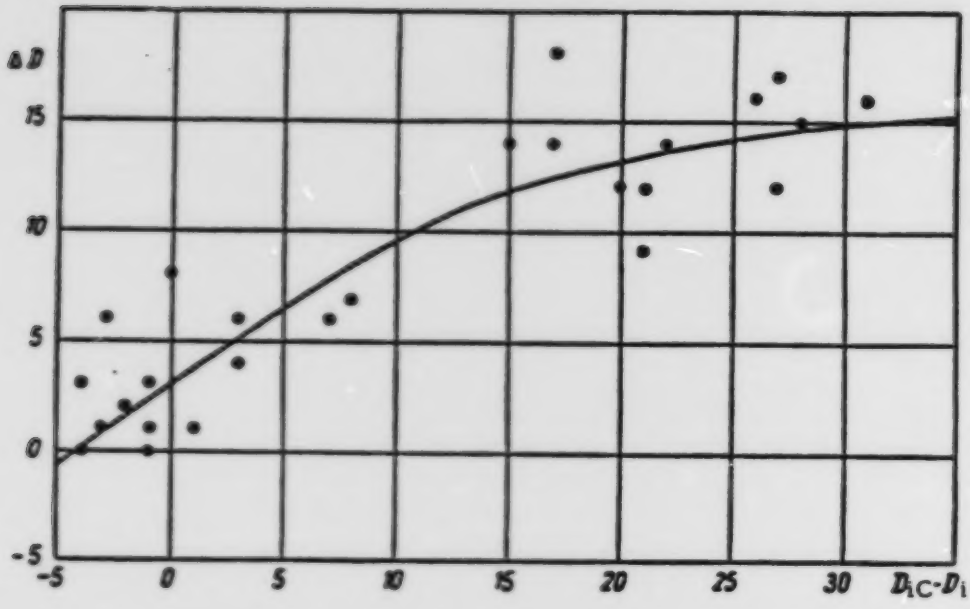
1- ΔH and $\sum q_0$ on the day of ice push; 2- ΔH and $\sum q_0$ on the preceding day.

Figure 18 Relationship between the stage increase required for breakup and the total heat input (after Shulyakovskii, 1963)

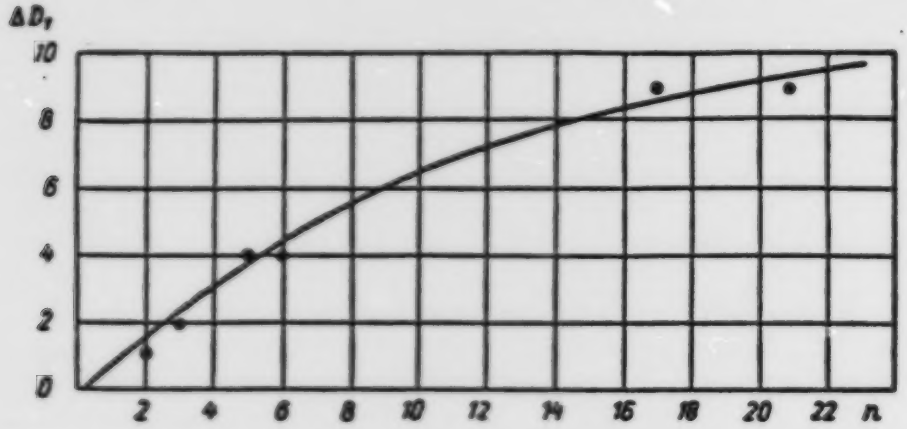


Correlation between early possible date of ice breakup on the Ob' River at Aleksandrovo, D', and the date of accumulation of $\Sigma\theta_+ = 30°$, D_{1B} at Barnaul.

Figure 19 Correlation between the date of breakup and the date on which 30°C-days of melt is accumulated (after Shulyakovskii, 1963)



Relationship between correction ΔD to the early possible date of ice breakup D' on the Ob' River at Aleksandrovo and the factor of rapidity of heat propagation over the basin $D_{1S} - D_{1B}$.



59. Variation of correction for cooling ΔD_1 with the number of days n of negative air temperature at Surgut.

Figure 20 Correction to the earliest possible breakup date on the basis of (a) the rapidity of heat propagation over the basin and (b) the number of days with negative air temperature during the melt period (after Shulyakovskii, 1963)

Various other techniques have been used to improve the reliability of the prediction techniques. For example, ΔY was correlated with the severity of the winter (°C-days of frost) and the degree days of melt in the spring prior to breakup (figure 21). However, no date was given to support the validity of this approach. Another attempt was to correlate the required ΣE to produce breakup at one particular location with the °C-days of frost of the preceding winter hence, the thickness of the ice cover (figure 22). This gives very poor results with an error of +300 °C-days over a total range of 1000°C-days. This could probably be improved however, by stratifying the data on the basis of the maximum winter stage.

Michel and Abdelnour (1975) defined breakup to occur when ice sheets would begin to detach from the upstream end of a solid ice sheet. Before detachment, the ice cover would submerge and oscillate due to the increasing discharge and flow velocity. Although this phenomenon occurs only under an idealized situation, they reasoned that its occurrence would be a good index at the time when the cover would become unstable and breakup would occur.

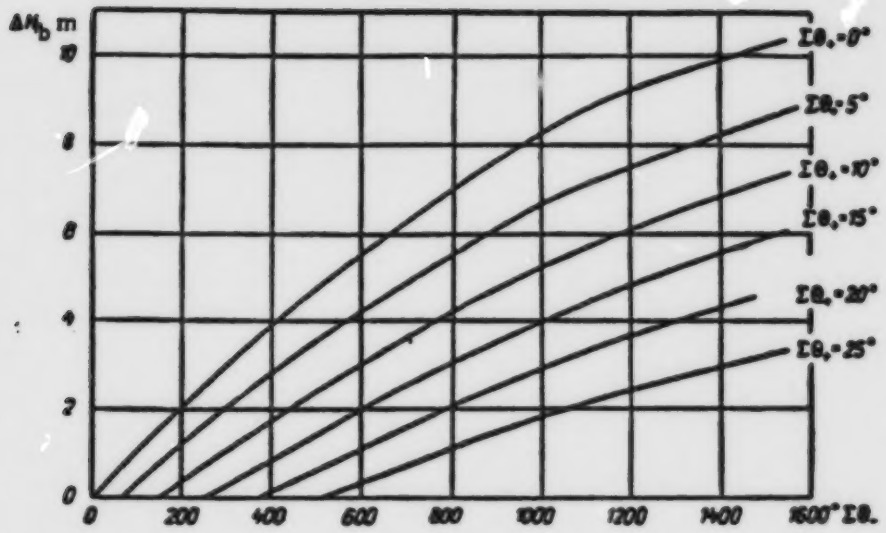
Dimensional analysis suggests that

$$[13] \quad f(v/\sqrt{\sigma/\rho},\ t_i/W,\ t_i/Y,\ \sigma/E) = 0$$

where v is the flow velocity under the cover; σ and E are the flexural strength and elastic modules of ice, respectively; t_i, Y and W are the thickness of ice, depth of flow and width of the stream, respectively, and ρ is the density of water. Laboratory tests were carried out in a flume using a model ice wax with a strength of 1/20 to 1/40 of real ice, and the results suggested that breakup should occur when

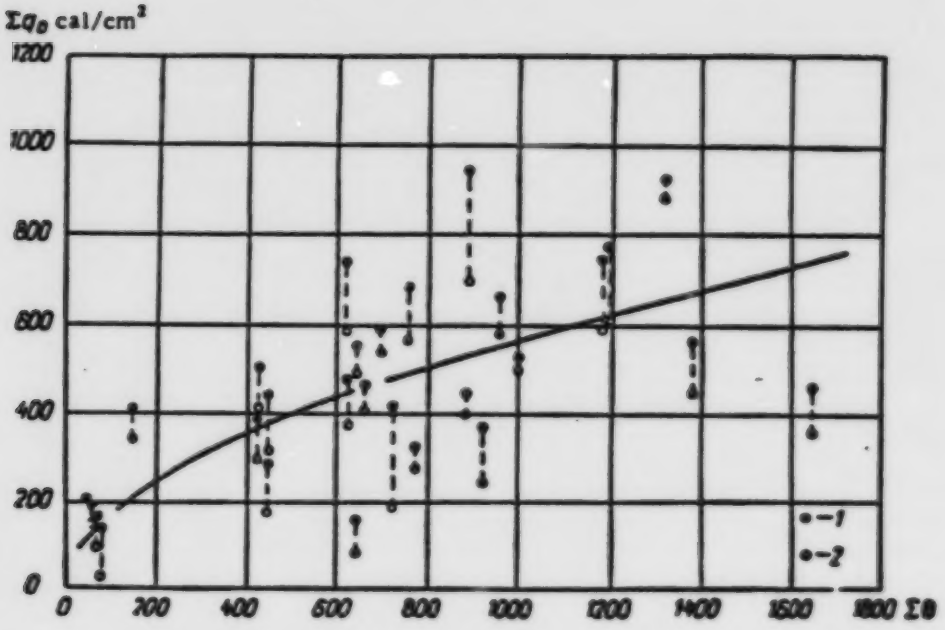
$$[14] \quad (v-v_o)/\sqrt{\sigma/\rho} = 0.055\ (t_i/W)^{3.8}$$

where $v_o = \sqrt{2g(\rho-\rho_i)t_i/\rho}$



Rise of water stage above the winter minimum, necessary for ice breakup on the Ob' River at the town of Kaluga, ΔH_w, as a function of the sum of the mean-diurnal negative air temperatures during the stable ice period, $\Sigma\theta_-$, for various values of the sum of the mean-diurnal positive air temperatures until ice breakup $\Sigma\theta_+$.

Figure 21 Stage increase necessary for breakup to occur for given degree days of melt as a function of the degree days of frost during the previous winter (after Shulyakovskii, 1963)



Amount of heat Σq_0, necessary for ice breakup on the Oskol River at the village of Ninovka, as a function of the sum $\Sigma\theta_-$ of mean-diurnal negative temperatures during the stable-ice period.

1- Σq_0 includes the value of q_0 during ice breakup day; 2- Σq_0 without the value of q_0 of ice breakup day.

Figure 22 Total heat input necessary to produce breakup as a function of the degree days of frost during the previous winter

and is the velocity at which floes can be entrained at the head of an ice cover. Thus, with a given ice strength and elastic modules, both which decrease over time as the ice temperature increases from atmospheric heating and an increasing velocity due to an increasing discharge from now melt runoff, one could forecast the onset of breakup.

The difficulty herein is that one must be able to define the ice strength at the time of breakup. This is extremely difficult and, ultimately, one must rely on some index of atmospheric heating ΣE, which when coupled with the increased discharge (which is directly related to the stage increase ΔY) gets this technology back to that proposed by Shulyakovskii. One other complicating factor makes this approach tenuous. The simplicity of the process does not allow for all the other types of deterioration patterns such as the development of mid-channel open leads, transverse cracking and the development of surges. These are all much more important and much more effective processes for destroying the integrity of an ice cover and would, therefore, override the idealized viewpoint adopted by Michel and Abdelnour.

Beltaos and Lane (1982) based much of their work on that of Shulyakovskii and demonstrated that a relationship between the rise required to initiate breakup, $Y_b - Y_f$ (where Y_b is the stage at breakup and Y_f is the maximum freeze-up stage) and the total heat input to the surface of the ice cover, ΣE could be used to predict the onset of breakup (figure 23) on the Nashwaak River. In this case, the total heat input was computed using Shulyakovskii's technique. It was apparent that the stage increase required to initiate breakup was significantly higher for low inputs of atmospheric energy (premature breakup) whilst for high energy input, a relatively small increase in stage was required. It should be noted that breakup was defined as the instant when a sustained movement of the ice cover begins. When reviewing WCS records, this was judged to occur when the stage suddenly drops or remains constant for a brief period in time following a rising trend. Thus, a decrease in stage results when the ice cover is set in motion, briefly reducing its resistance to the flow.

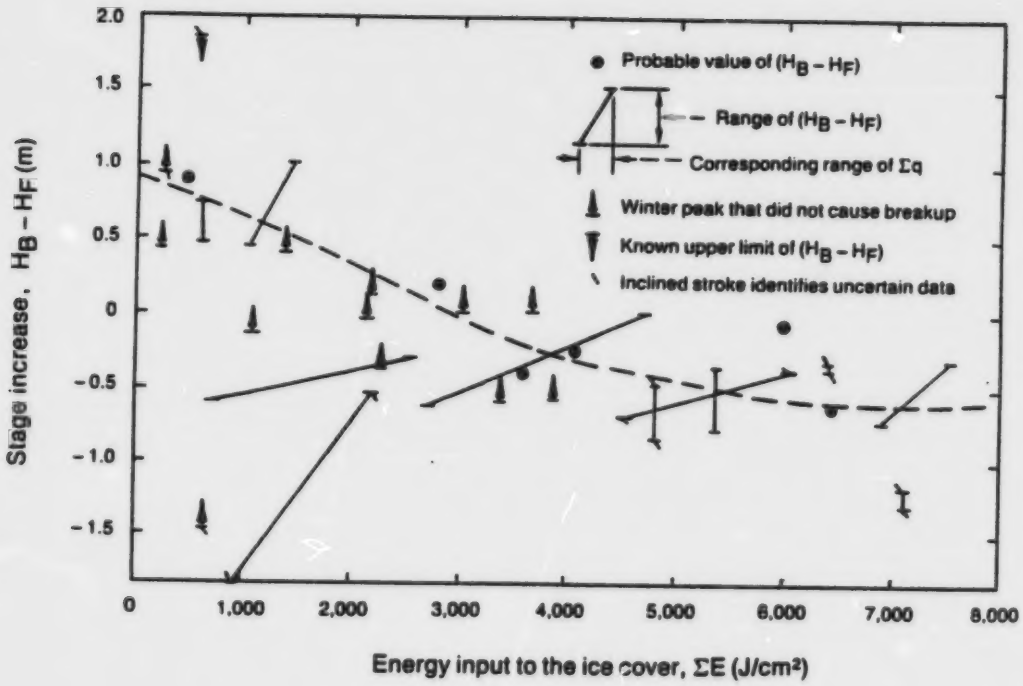


Figure 23 Total heat necessary to produce breakup as a function of the stage increase above the freeze-up stage (after Beltaos and Lane, 1982)

The data in figure 23 shows considerable scatter and certainly it must be used with considerable judgement. One of the difficulties is rationalizing the effect of the ΣE term. The onset of breakup can be related to three processes, each due to a different combination of meteorologic, hydrologic and hydraulic conditions. For discharge related breakup, the movement of the ice cover is related only to the increase in stage due to an increase in discharge from melt upstream in the basin. When the stage increases above Y_f by some amount (maybe t_i), the cover moves downstream and either breaks or remains solid. In either case, there is some perturbation to the gauge height and breakup is assumed. In this situation, the ΣE term enters into the calculation as an index of the discharge, and the prediction of breakup can be made by determining the amount of melt required to produce an increase in stage sufficient to raise the ice cover above the freeze-up stage. Provided there is sufficient snow to generate enough runoff, breakup should be initiated.

In this situation, the ΣE term is primarily responsible for a change in discharge and its effect on the strength of the cover is minimal at low ΣE values. In fact, in figure 23, it appears that when ΣE is low and yet breakup occurs, Y_b-Y_f is poorly correlated with ΣE. This suggests that this approach is not a good way to predict breakup because it does not provide a way to handle the stage increases which results from the development of surges which can occur on a local scale.

In a thermal breakup situation, there is insufficient snow on the ground to enable ΣE to produce enough runoff to lift the cover. The ice deteriorates gradually and eventually breaks up. In this instance, ΣE is an index of the strength of the cover. This type of breakup requires considerable time to evolve and as illustrated in figure 23, Y_b-Y_f does not have to be large to produce breakup. Forecasting this condition, even though the process is considerably simpler than for the discharge, related breakup is very difficult because the initial ice condition can be so variable.

The other type of breakup which can occur is the mechanical breakup. In this situation, the release of an ice jam upstream and the destruction

of the ice cover by the surge of ice and water occurs very rapidly without the necessary ΣE to deteriorate the ice cover and even without significant runoff being generated. In this situation, the breakup of the cover occurs mostly by chance, depending on the availability of ice upstream and its ability to get to the location of interest. Historically, this can occur with hardly any melt in the basin (or at the site of interest) or after more melt than would be expected if the ice jam upstream stabilized and could not get to the site of interest. Therefore, the approaches by Shulyakovskii and Beltaos do not look at all the physical processes in the river, but rather attempt to correlate breakup of one particular location with some rather global type of parameters assuming a very narrow set of idealized processes.

Beltaos (1984,85) looks at breakup in the broader context of a whole river reach. He indicates that a necessary condition to breakup is that $Y>Y_f$ over a significant portion of the reach. As the stage increases, longitudinal cracks or fractures form parallel to the bank. On wide rivers, the cracks form in pairs, one on each bank, and the cover floats free. For narrow channels or for very thick ice covers, one crack along the channel centre line may occur if λW is less about 4, where W is the channel width and

$$[15] \quad \lambda = \sqrt[4]{12\sigma/4E\ t_i^3}$$

in which all the terms have been previously defined.

After the longitudinal cracks develop, transverse cracks also form, most probably from horizontal bending caused by differential stresses due to meandering planform of the river. This produces broken sheets which are in the order of 4-5 W in length and about 1/2 W in width. As the stage increases, the width increases, and the sheets get more room to move, eventually accumulating against some naturally occurring impediment. Once the stage increase is sufficient to move all the ice sheets downstream, the critical ice clearing discharge has been exceeded, and breakup has been established.

In conceptual terms, the average length of the ice sheet, l_i, can be written as

$$[16] \quad l_i = f\,(t_i, \sigma, \tau, W_i, \Phi_i)$$

where τ is the driving force moving the ice sheet (shear stress, for example) and Φ_i is the central angle of a large curved river. The river width necessary to just pass an ice sheet of dimension l_i can be defined as

$$[17] \quad W_b = f_1\,(W_i, l_i, L, \Phi)$$

where L and Θ are typical length and curvature scales of the channel and

$$[18] \quad W_b/W_i = f_2(l_i/W_i, L/W_i, \Phi)$$

Because the last two terms in the brackets of equation [18] are constant from year to year for any given river and introducing equation [16]

$$[19] \quad W_b/W_i = f_3(t_i/W_i, \sigma/\tau)$$

To produce an instability such that the large ice sheets can move downstream, the stage and the width must increase beyond the freeze-up stage and width. Thus,

$$[20] \quad W_b/W_f = f_4(t_i/W_f, \sigma/\tau)$$

and because the width is a function of the stage, then

$$[21] \quad Y_b/Y_f = f_5(t_i/W_f, \sigma/\tau)$$

where Y_b is the stage necessary to clear the channel of ice or produce breakup. Once Y_b is determined, then the necessary discharge to produce breakup can be computed by

$$[22] \quad Q_b = (0.62/n_o)[Y_b-(\rho_i/\rho)t_i]^{5/3}S^{1/2}$$

where n_o is the composite roughness and S is the river slope.

This suggests that there is an upper limit on the stage as determined by the requirements of an ice clearing discharge, which, if exceeded, will produce breakup and cause running ice to be evident. As in the previous discussion, both t_i and τ depend on ΣE because t_i is reduced as ΣE increases and snowmelt increases the flows in the river.

Figure 24 illustrates some typical values of the ice clearing discharge as plotted against t_i/W_f. There appears to be a rather well defined upper limit, obviously related to situations where very little ice deterioration occurred and mechanical breakup prevailed. It is apparent, however, that from year to year both t_i and W_f are generally constant, and therefore, the channel clearing discharge becomes much more dependent on the strength of the ice cover and the shear on the bottom of the ice. Also, the data is from a variety of very dissimilar rivers, and no attempt has been made to include the L/W_f and Φ terms to differentiate the morphological characteristics. The problem still is to explain the scatter of up to 50% of the observed value by the appropriate use of ΣE to reduce the strength of the cover.

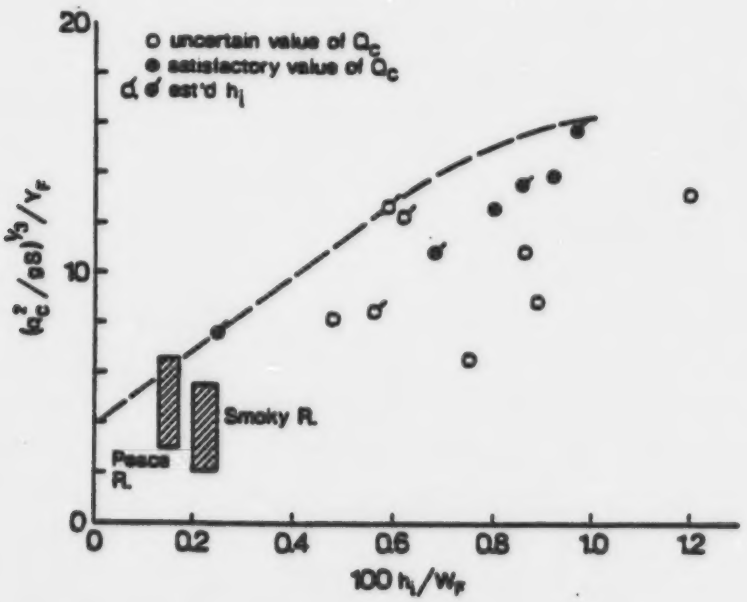


Figure 24 Dimensionless ice-clearing discharge versus the dimensionless ice thickness

All of the current models of breakup suffer the disadvantage of being too site specific. This has resulted primarily from the type of data readily available for analysis. The difficulty this presents is there is no systematic look at reach average conditions. The instabilities and subsequent breakup which is being described could occur in isolation of any location without being generalized over a sufficiently long reach. The other difficulty is that the models all lack a physical basis to describe the process of ice deterioration due to the meteorological conditions. Instead, the relationship between the ice strength and ΣE is used and there is no separation between the affects of ΣE on the deterioration of the ice cover and its effect on the generation of runoff.

4.2 Model Development

4.2.1 Conceptualization

Observations on the Athabasca River (section 3.1) suggest that breakup on a reach scale occurs by a combination of mechanical and thermal processes. Prior to the occurrence of snowmelt, the ice cover is relatively intact. However, the natural variability of the freeze-up process results in a very heterogeneous ice cover. The low discharges at which the cover forms and the subsequent low winter flows result in an ice cover situated at extremely low stage in the portion of the channel where changes in width are most sensitive to changes in stage. Once snowmelt begins, the flow increases and the ice cover is lifted free of the banks against which it is floating or the shallow areas on which it is resting. As the stage increases, the top width increases, but with the constant ice width, the net effect is to produce open water. The amount of open water is a function of both the initial stage (i.e., stage prior to the generation of runoff) and the channel characteristics.

The lower albedo of the water, compared to that of ice, results in the creation of local heat sinks which produce larger local areas of open water. This then accelerates the ice melting process and leads to ever increasing proportions of open water. Concurrently, once the snow cover has melted the ice cover begins to absorb energy and thin out. This

thinning of the ice cover also accelerates the thermal process by which open water is generated because there is a decreasing amount of ice for the water to melt.

This process leads to the overall weakening of the reach averaged ice condition. The combination of a thinner ice cover and greater fraction of open water, due to both the melting of the ice cover and the increased width as the stage increases, leads to an unstable situation. This facilitates the mechanical destruction of the ice cover, the formation of ice jams and, subsequently, the production of surges which ultimately destroy sections of competent ice and produce open water conditions. The model evaluated herein is meant to describe the process by which an ice cover goes from being totally intact to the point of incipient instability when ice jams form, surges develop and a generalized ice run occurs.

4.2.2 Formulation

Consider an elemental area of an ice cover (figure 25) having an area $\Delta x \Delta Z$ of which Θ is the fraction of open water. For an air temperature T_a and solar radiation H, the net heat transfer to the water surface be given by

$$[23] \quad \Theta \Delta x \Delta Z \left[\Gamma(1-\alpha_w)H + h_w T_a\right]$$

where the first term is the radiation component and the last term is the convective component and where α_w is the albedo of water; Γ is the exposure factor which reduces the effective solar radiation; H is the average daily radiation flux, h_w is the convective heat transfer coefficient above water; and T_a is the average daily air temperature.

Additional open water, Θ is also generated when the ice cover is lifted free of its pre-melt position by the increase in discharge. Assume that the width, W, can be written as a function of stage, Y,

$$[24] \quad W = mY^n$$

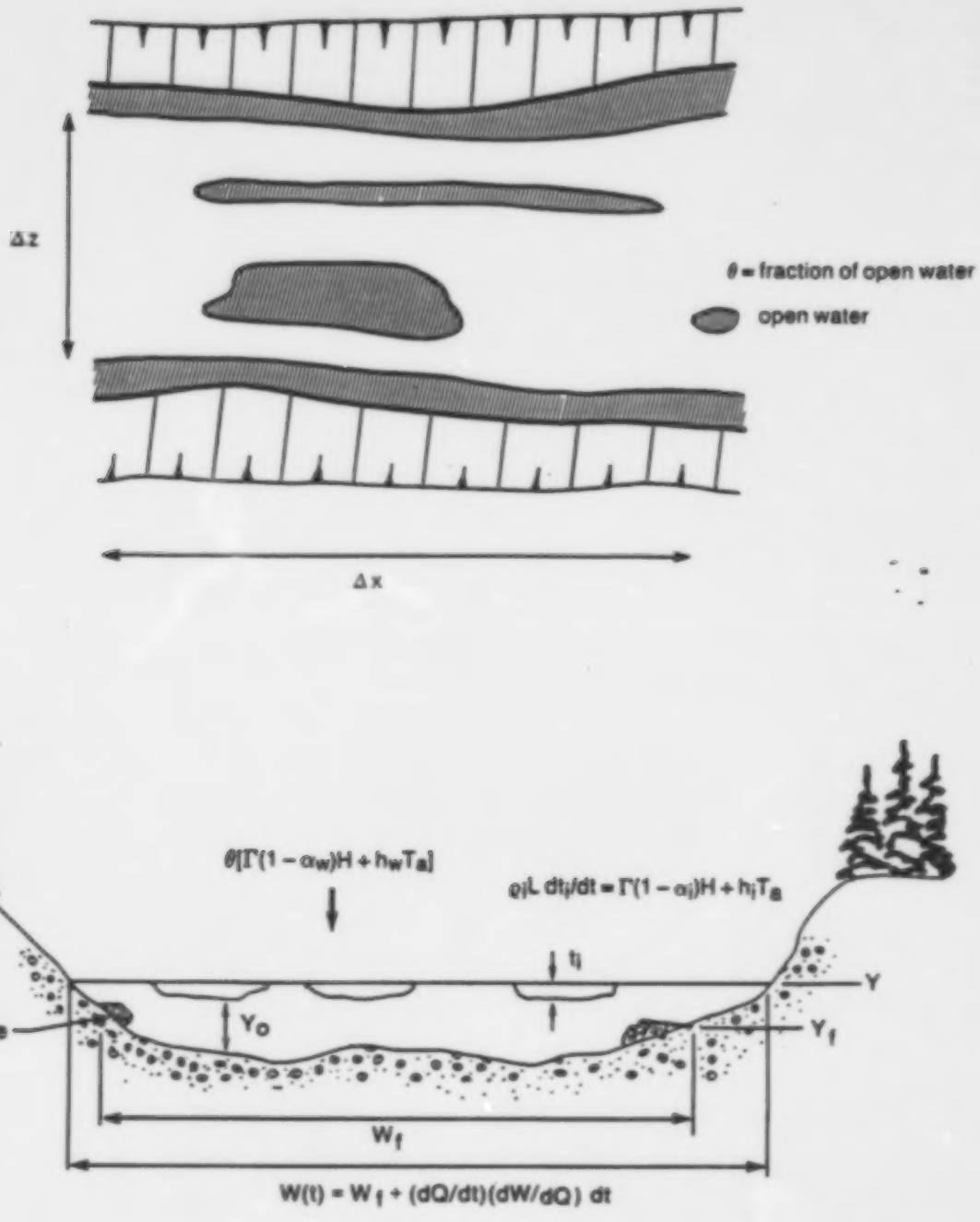


Figure 25 Schematic representation of the processes which cause the development of open water

and the stage is a function of the discharge, Q

$$[25] \quad Y = pQ^q$$

then

$$[26] \quad W = mp^n Q^{nq}$$

where m, p, n and q are constants derived from the channel characteristics and the roughness of the ice cover. Thus, if W_f is the width of the river stage just prior to breakup, then the effective open water, as a fraction of the initial width, can be designated as

$$[27] \quad \Theta = (W/W_f) - 1$$

To formulate the model, one must describe the process by which open water is created. In this situation, a number of assumptions must be made.

1. Open water is initially initiated by the rise in stage due to the increased runoff due to snowmelt and is given by equation [27].

2. All the energy that is absorbed by the open water leads to the creation of the equivalent amount of additional open water the next day. Thus

$$[28] \quad d\Theta/dt = \Theta\,[\Gamma(1-\alpha_w)H + h_w T_a]/(\rho_i L t_i)$$

3. Ice thinning begins on the day when all the snow on the ice cover has melted and occurs at a rate at which energy is absorbed into the ice cover,

$$[29] \quad \Gamma(1-\alpha_i)H + h_i T_a$$

where α_i is the albedo of the ice cover, h_i is the convective heat transfer coefficient above the ice surface and the other terms have been

defined previously. Additional energy is added to elemental ice cover by friction caused by the flow under the ice cover. This can be defined as the stream power, $\tau_o v_i$ exerted by the flow on the bottom of the ice cover. The flow Q, can be defined as

$$[30] \quad Q = (1/n_o)\ (Y_o/2)^{2/3}\ YWS^{1/2}$$

where n_o is the composite roughness and Y_o is the depth of flow under the ice from which the velocity, v_i, affecting the ice cover can be defined as

$$[31] \quad v_i = (1/n_o)\ (Y_o/2)^{2/3} S^{1/2}$$

if one assumes that the roughness of the ice cover and the bed are equal. This is not an unrealistic assumption during the pre-breakup period, because the heat transfer of the ice cover tends to produce ripples on the underside of cover and, hence, increase its roughness substantially. The shear stress τ_i on the ice cover can be computed from

$$[32] \quad \tau_i = \rho g Y_i S$$

where Y_i is the flow depth associated with the ice cover and is 0.5 Y_o because of the assumption of equal roughnesses. Thus

$$[33] \quad \tau_o v_i = 0.5\ \rho g Q S$$

and the overall change in the ice thickness can be written as

$$[34] \quad dt_i/dt = (-1/\rho_i L)\ [\Gamma(1-\alpha_i)H + h_i T_a + 0.50\ \rho g Q s]$$

Equations [27], [28] and [34] can be combined and solved analytically. However, the radiation, air temperature and discharge are all given in discrete daily units and, therefore, it is more practical to write the equations in a discrete form. Assuming

$$[35] \quad d\Theta = \Theta_{j+1} - \Theta_{j+1}$$

$$[36] \quad dt = t_{j+1} - t_j = \Delta t = 1$$

then

$$[37] \quad \Theta_{j+1} = \Theta_j \, (W_j/W_f) \, \{1+[\Gamma(1-\alpha_w)H + h_w T_a]/(t_{j+1}\rho_i L)$$

where

$$[38] \quad t_{j+1} = t_j - (1/\rho_i L) \, [\Gamma(1-\alpha_i)H + h_i T_a + 0.5 \, \rho g Q_j S]$$

and

$$[39] \quad W_j/W_f = mp^n \, Q_j^{nq}/W_f$$

By defining, from the records, the maximum thickness of the ice cover, the day on which the stage begins to systematically increase and the day when the last of the snow has melted, one can do the calculation to compute the increase in the fraction of open water on the basis of the daily values of discharge, solar radiation and air temperature.

4.2.3 Calibration

Calibration and/or application of the model requires the evaluation or definition of seven variables, aside from those variables associated with the channel characteristics. In general, the albedos and the heat transfer coefficients should be universal and not a function of location or time. Hence, they can be assigned with a reasonable degree of confidence from other studies. The initial ice thickness and the start dates on which the model computations begin, are assigned from observations of river stages and snowpack characteristics which are specific to the year being analyzed. The only variable requiring calibration is the exposure factor, Γ, which is a function of the channel planform, valley wall characteristics and the sun angle.

Albedo

The surface albedo is a function of the solar angle and characteristics of the surface. Anderson (1954) summarizes some investigations which evaluated the albedo of water. He suggests that solar radiation is reflected from both the water surface and a stratum of relatively opaque

water composed of bubbles and suspended material located just below the surface. The reflectivity also tends to increase when the surface is hydrodynamically rough and under cloudy conditions, however, it is never more than 0.09 for solar angles greater than 30° and under overcast skies, it averages about 0.08. Measurements in northeastern USA suggest α_w varies between 0.034 and 0.097, while measurements in California suggest α_w as high as 0.105. Anderson found that for a sun angle of 30°, which is about average for the Athabasca River during the spring period, the albedo varies between 0.05 and 0.12, but typically has values of about 0.075 for a variety of cloud conditions from scattered to broken, to completely overcast.

The albedo of ice is not as well documented as that of water. Budyko (1956) suggests that the albedo of ice varies from between 0.80-0.95 for very white ice covered by new snow, to 0.40-0.50 for "polluted" ice. Bolsenga (1977) found that the albedo of ice varied substantially with both the solar angle and the quality of the ice. However, under decaying ice conditions, the albedo varied from 0.45 at low sun angles to 0.20 at high sun angles. At a sun angle of 30°, he estimated the albedo to be 0.35. In another study, Bolsenga (1969) found that snow free Great Lakes ice had albedos ranging from 0.10 to 0.46 for a variety of solar angles and cloud conditions, but on the whole, averaging 0.31.

Convective Heat Transfer Coefficients

The convective heat transfer coefficient is not so well defined as the albedo because it used to define a much more complicated process. The convective heat transfer coefficient, h, is often used to describe all the temperature-related heat transfer processes in one lumped coefficient. Unfortunately, many investigators, Ashton (1983) for example, also include the effects of the solar radiation in their definition of h. Thus, they probably over-estimate its true value.

Andres (1984), for the Upper Peace River, found that after separating out the effects of solar radiation, the convective heat transfer coefficient above water, h_w, was in the order of 15 W/m^2-°C. Ashton

(1983), in analyzing the melting of lake ice, found that h_i varied between 10-20 W/m^2-°C, depending on the year and the time of year. It is possible that the higher values that he calculated include a significant contribution from solar radiation. Therefore, the lowest value of h_i probably defines the time when there was limited solar radiation and, therefore, would be most consistent with the definition used herein. Thus, h_i should be in the order of 10 W/m^2-°C.

Channel Geometry

Limited data exists for the reach of the Athabasca River between the mouth of the Pembina River and Fort McMurray. For convenience, the channel characteristics of Athabasca were considered as being typical for the entire reach. The channel width was found to correlate with gauge height, Y, to give

$$W = 310Y^{0.15}$$

The relationship between the gauge height and the discharge was not so well described and, somewhat, a function of the procedure used by WSC to define the discharge during the pre-breakup period. Nevertheless, correlating the WSC estimated mean daily discharge with the mean daily stage produced the relationship

$$Y = 1 + 0.00002\ Q^{1.95}$$

and, therefore,

$$W/W_f = [(1 + 0.00002\ Q^{1.95})/(1 + 0.00002\ Q_f{}^{1.95})]^{0.15}$$

where Q_f is the base discharge prior to the occurrence of snowmelt runoff.

Initial Conditions

The most difficult initial condition to express adequately is the maximum winter ice thickness. WSC records indicate that the maximum ice

thickness varies between 0.75 and 1.0 m at Athabasca. This is a relatively wide range and must be substantiated by ice thickness estimates for the year of interest. Measurements at Fort McMurray (Malcovish and Andres, 1988) in March 1986, revealed a mean ice thickness of 1.1 m, whilst similar measurements at Hondo (Van Der Vinne, 1988) showed a mean ice thickness of 0.90 m. Therefore, the starting or maximum ice thickness adopted in the analysis was 1.0 m. This value seems to be relatively typical from location to location. However, there is no doubt, that much thicker ice is evident in areas where significant frazil accumulations occur. This thick ice is probably offset by thinner ice along thalwegs, but, undoubtedly, is the last remaining ice to be removed by the surges.

The start date for the computation of the open water development was judged to be on March 16. This day marked the end of a relatively long period of more or less constant stage. After March 16, there was a steady increase in stage.

The date on which the ice began to thin, assumed to be the last day of substantial snow cover on the ice, is somewhat more difficult to objectively define. Different locations of a basin respond differently to snowmelt and redistribution. Some locations accumulate snow, while others lose their snow before the snowpack is completely melted. The surface of the ice is no exception. There is, no doubt, a redistribution of snow as well as a net reduction due to melting. For the purposes of this study, it was assumed that the snowpack on the ice behaved similarly to that of the basin, and that the basin snow condition was typified by the response of the snow pillow at Twin Lakes. From figure 2, it shows a negligible snow cover in the basin first occurred on April 4. Given the probability of a reduced snowpack on the ice surface due to its exposed location, it was assumed that the snow was essentially gone from the ice surface one week prior to that at the snow pillow. Thus, a net thinning or melting of the ice on the Athabasca River first began on March 28. This date may appear to be early, however, given the exposed nature of the ice cover, it was arbitrarily chosen, reflecting a somewhat earlier depletion of the snow cover in exposed locations.

The adopted values of the coefficients described above are summarized in table 4.

Table 4 Summary of model parameters

Parameter	Value
Ice albedo, α_i	0.4
Water albedo, α_w	0.05
Convective heat transfer coefficient over water, h_w (W/m^2-°C)	15
Convective heat transfer coefficient over ice, h_i (W/m^2-°C)	10
Starting ice thickness, t_o (m)	10
Date of first stage increase	March 17
Date of last snow	March 28

The exposure factor, Γ, was evaluated by applying the model to the data summarized in table 4 with daily values of air temperature, solar radiation and discharges from figures 3 and 6 as inputs. Figure 26 compares the calculated values of the fraction of open water to those actually measured. It appears that the most appropriate value of Γ for this reach of river is 0.75. The goodness of fit is quite remarkable, except for the April 12 point. It may be possible that the April 12 value was over-estimated, given the inherent inaccuracies of estimating the fraction of open water. However, the rate of change is quite well represented over the entire breakup period and given the simplicity of the model, its results appear to be sufficiently accurate.

The shape of the curve defining the fraction on open water is a function of two processes. Initially, prior to about April 8, the open water is primarily a function of the discharge and because the ice is still relatively thick and there is less than 5 % open water, variations in the thermal processes do not contribute substantially to changes. However, after April 8, the stage is relatively constant and with a substantial fraction of open water now apparent, the shape of the curve is primarily a function of the heat transfer processes. The incremental

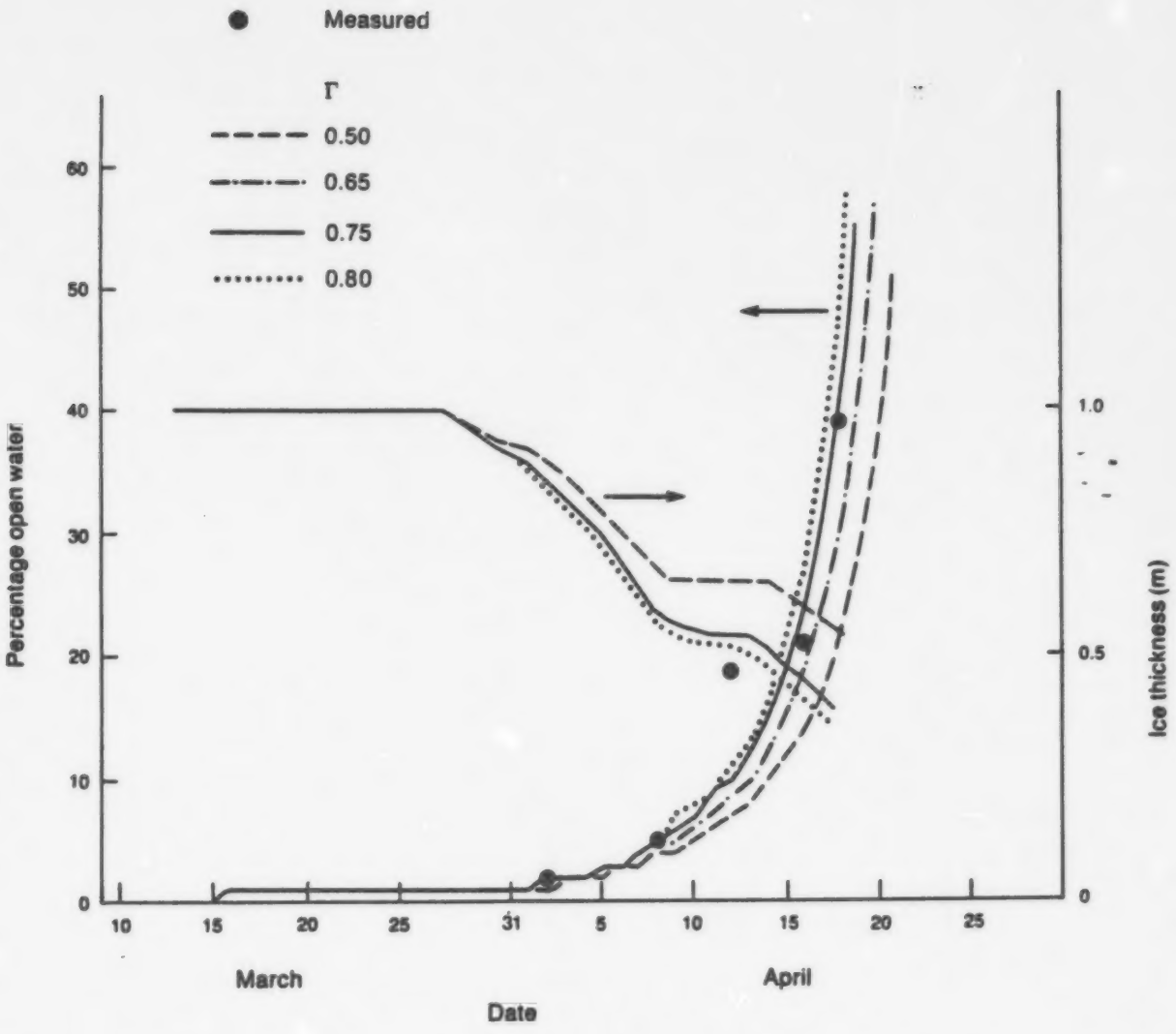


Figure 26 Calibration of the exposure factor and sensitivity of the ice thickness reduction to variations in the exposure factor

increase in the fraction of open water reflects the warmer temperatures, the greater solar radiation and the thinner ice cover.

4.2.4 Sensitivity Analysis

A sensitivity analysis was carried out to evaluate the impact that errors in some of the more subjectively prescribed values of the model parameters would have on the model output. Each of the six parameters, including the initial ice thickness, the albedos of ice and water, the convective heat transfer coefficient for ice and water and the exposure factor were altered within a prescribed range and compared against the field data and the output from the adopted calibration run. It should be noted that the channel characteristics, the first day of discharge increase and the last day of snowmelt were not evaluated. These can generally be prescribed for any given year in a relatively objective manner.

The exposure factor was varied between 0.5 and 0.8 and the results indicate that the model output is quite sensitive to its prescribed value. It is apparent that significant deviation occurs once the exposure factor is reduced below 0.65 (figure 26) and at 0.5, the time of breakup would be delayed by about 3 days. Increasing the exposure factor to 0.8 does not alter the model output significantly, but it accelerates the breakup process as would be expected. Also, as anticipated, most of the variation occurs following April 8, at which point heat transfer, rather than a stage increase, becomes the dominant reason for open water generation.

Figure 26 also indicates the impact variations in the shading factor have on the computed ice thickness. Unfortunately, no measurements of thickness in the pre-breakup period are available, however, observations at Hondo (Van Der Vinne, 1988), suggest that ice thicknesses at breakup are generally 0.5 m and greater. This suggests that the ice thickness computations are not unrealistic and exposure factors between 0.5 and 0.80 are appropriate. It is apparent from this figure that the increased ice thickness caused by a reduction in the shading factor may contribute to the delaying of breakup.

Variations in the albedo of water between 0.3 and 0.5 (figure 27) and the albedo of ice between 0.025 and 0.10 (figure 28) do not seem to make a significant difference in the output, and the development of open water was satisfactorily represented within ±1 day for the entire range of values chosen for each parameter. Similarly, variations in the convective heat transfer coefficients for water from 10 W/m^2-°C to 25 W/m^2-°C and for ice from 5 W/m^2-°C to 20 W/m^2-°C do not produce any striking differences from the calibration curve (figures 29 and 30).

It is interesting that figures 28 and 30 indicate almost no change in the production of open water due to changing either the albedo or the heat transfer coefficient of the ice cover even when these parameters significantly change the computed ice thickness. This suggests that the variation in the rate of open water development are much more sensitive to the amount of solar radiation and sensible heat reaching the water surface and the subsequent melting out the ice cover than to the effects of a varying ice thickness.

The final variable tested was initial ice thickness. It was apparent (figure 31) that variations in the adopted initial ice thickness created significant changes to the model output. If the minimum ice thickness (0.75 m) was assumed, the onset of breakup was accelerated by three days. If the thickest ice cover (1.2 m) was assumed, breakup would be delayed by slightly more than one day. Therefore, the natural variation in the ice thickness can significantly affect the time of breakup.

In addition to the above, the impact on the production of open water and, subsequently, the date of breakup of some hypothetical conditions could be assessed for the given climatic conditions which prevailed in March and April 1986. For example, maximizing the open water production by assuming the most optimum conditions for a quick breakup (i.e., minimum initial ice thickness, maximum energy transfer rates) leads to incipient breakup some nine days before it would have been observed (figure 32). Minimizing the rate at which open water would be generated by assuming a thick ice cover and low efficiency in energy transfer resulted in breakup being delayed by four or five days. This suggests

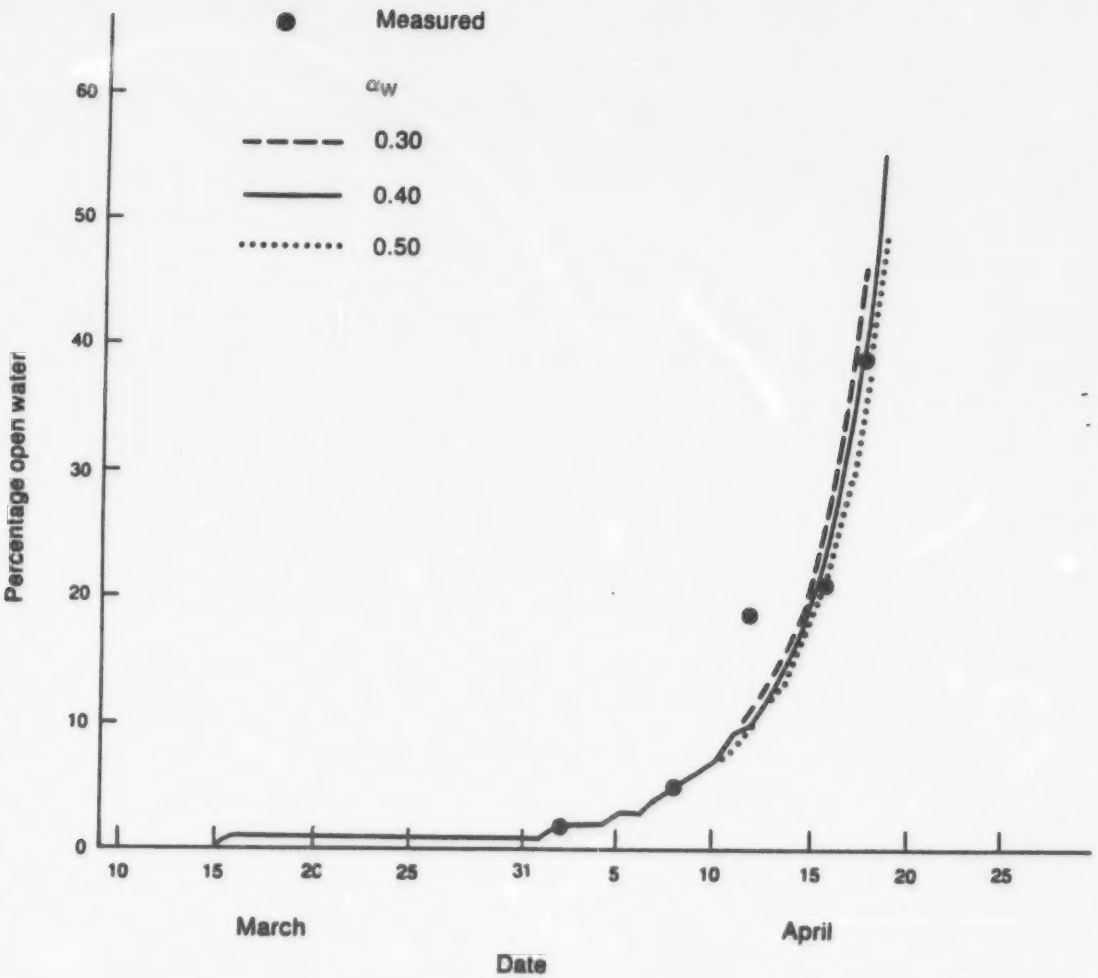


Figure 27 Sensitivity of the rate of open water generation to variations in the albedo of the water surface

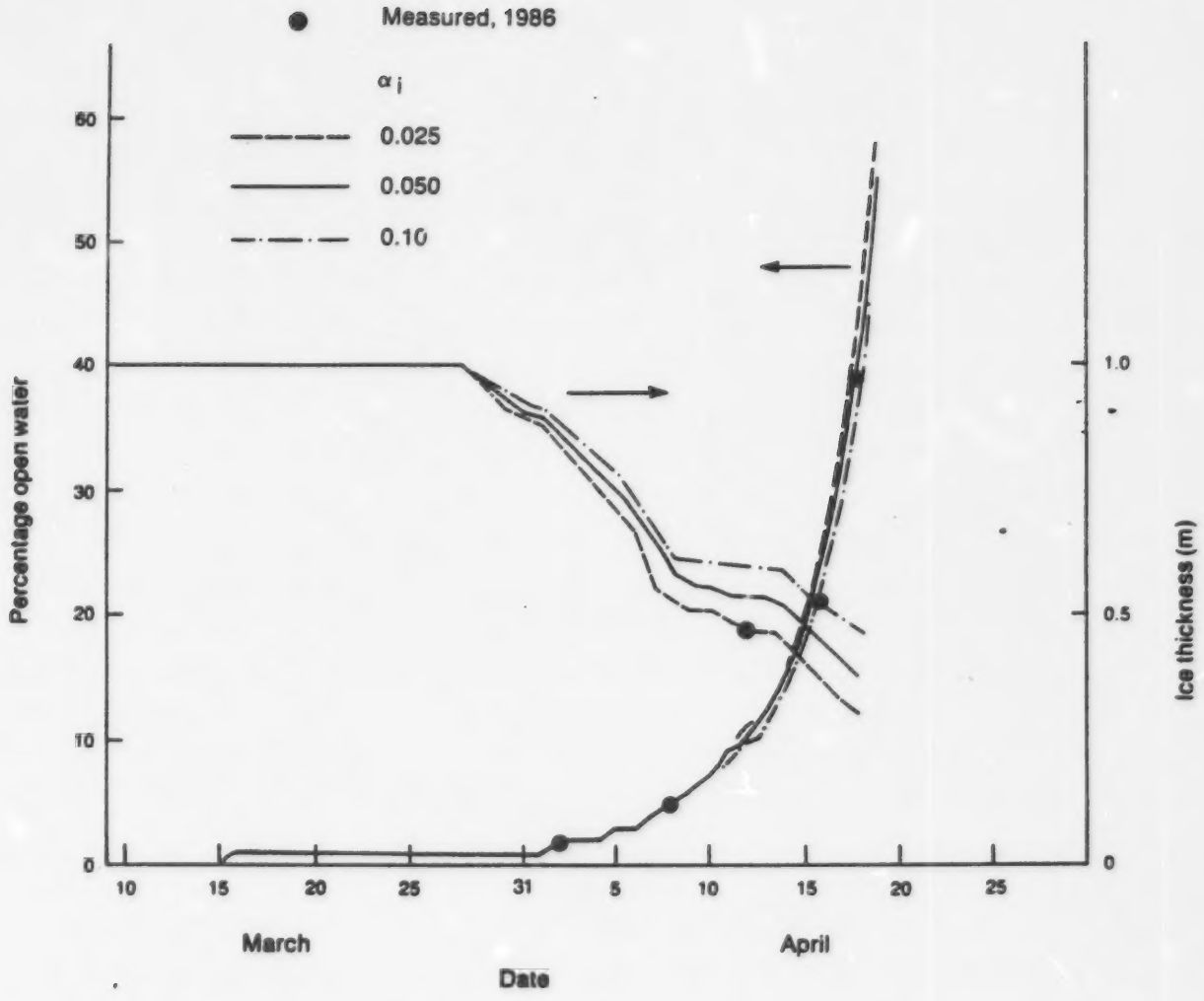


Figure 28 Sensitivity of the rate of open water generation and ice thickness reduction to variations in the albedo of the ice cover.

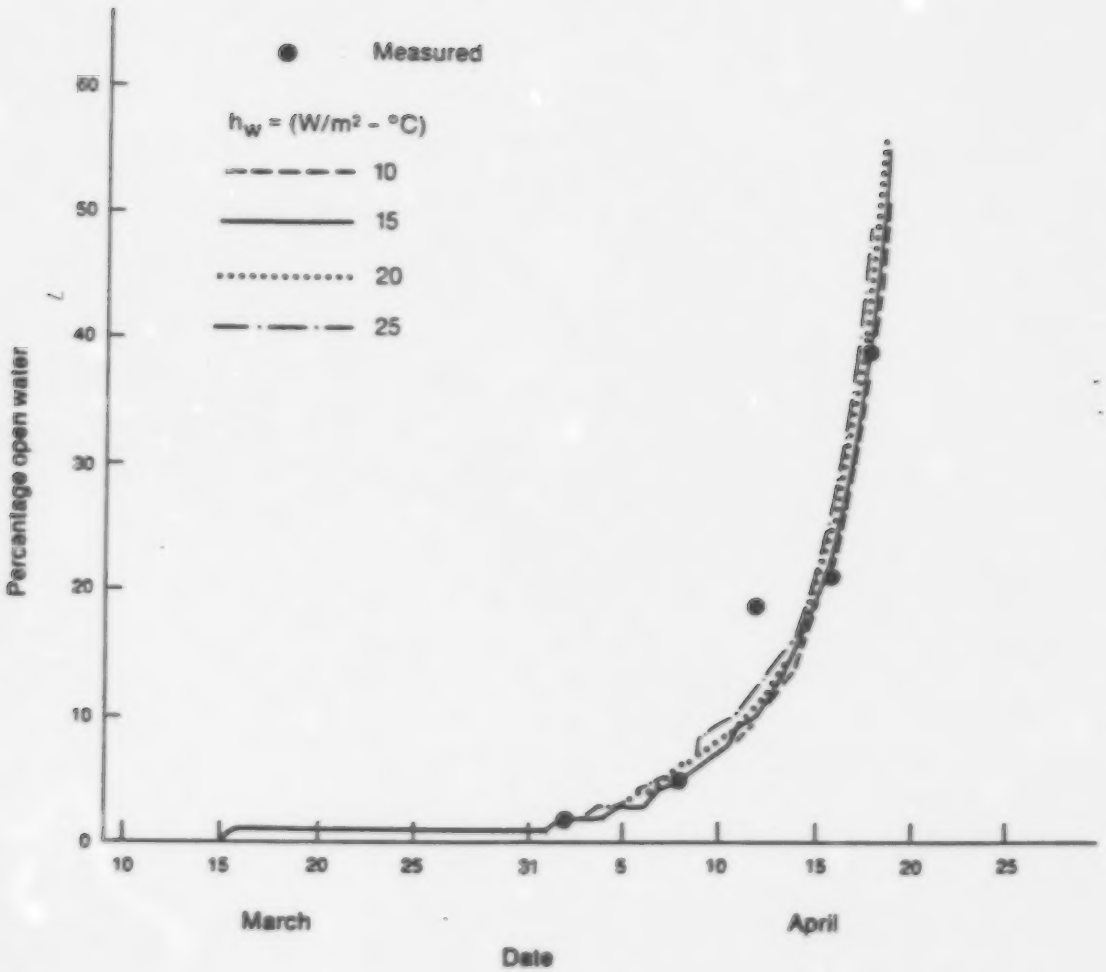


Figure 29 Sensitivity of the rate of open water generation to variations in the heat transfer coefficient over water.

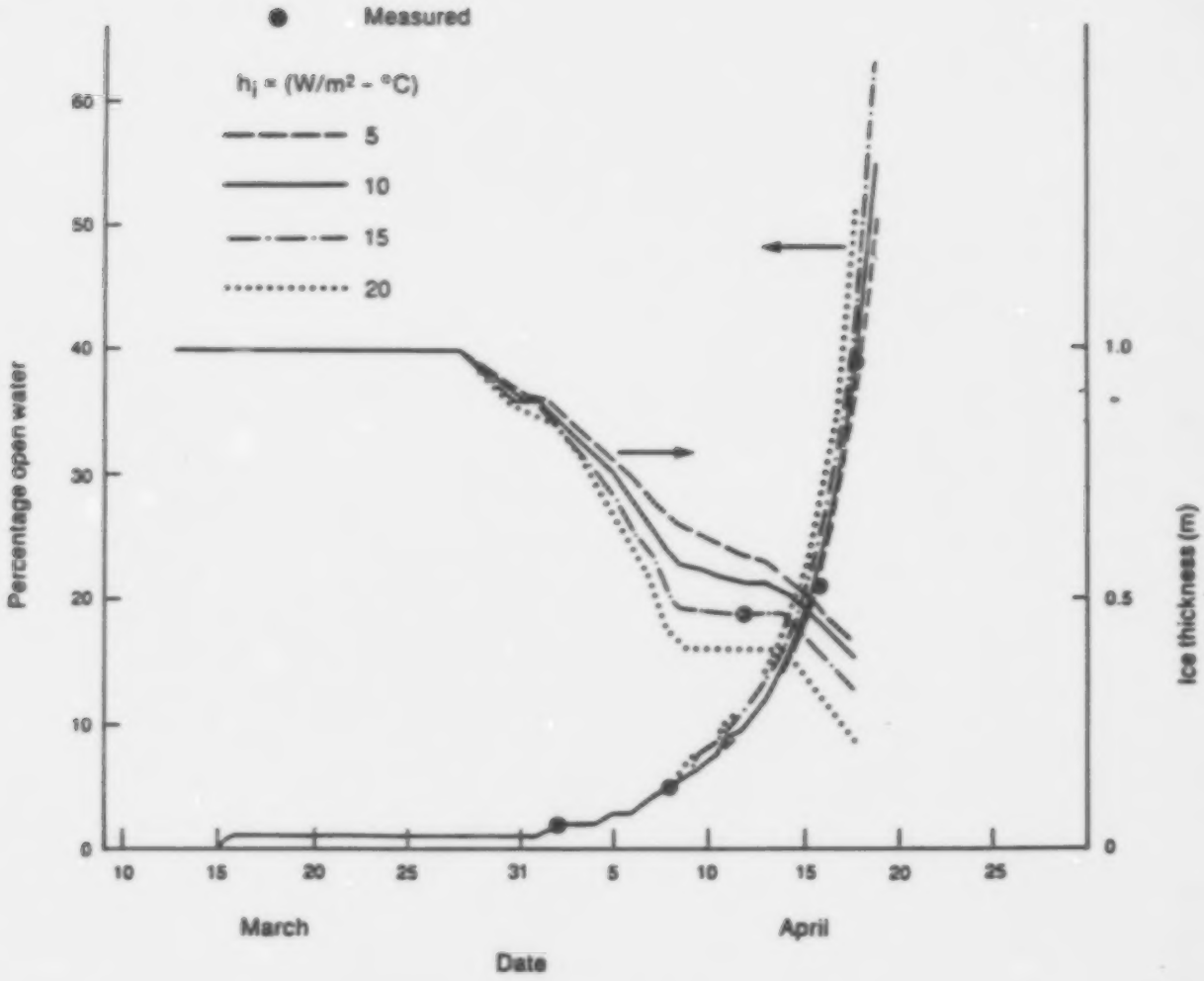


Figure 30 Sensitivity of the rate of open water generation and of ice thinning to variations in the heat transfer coefficient over ice.

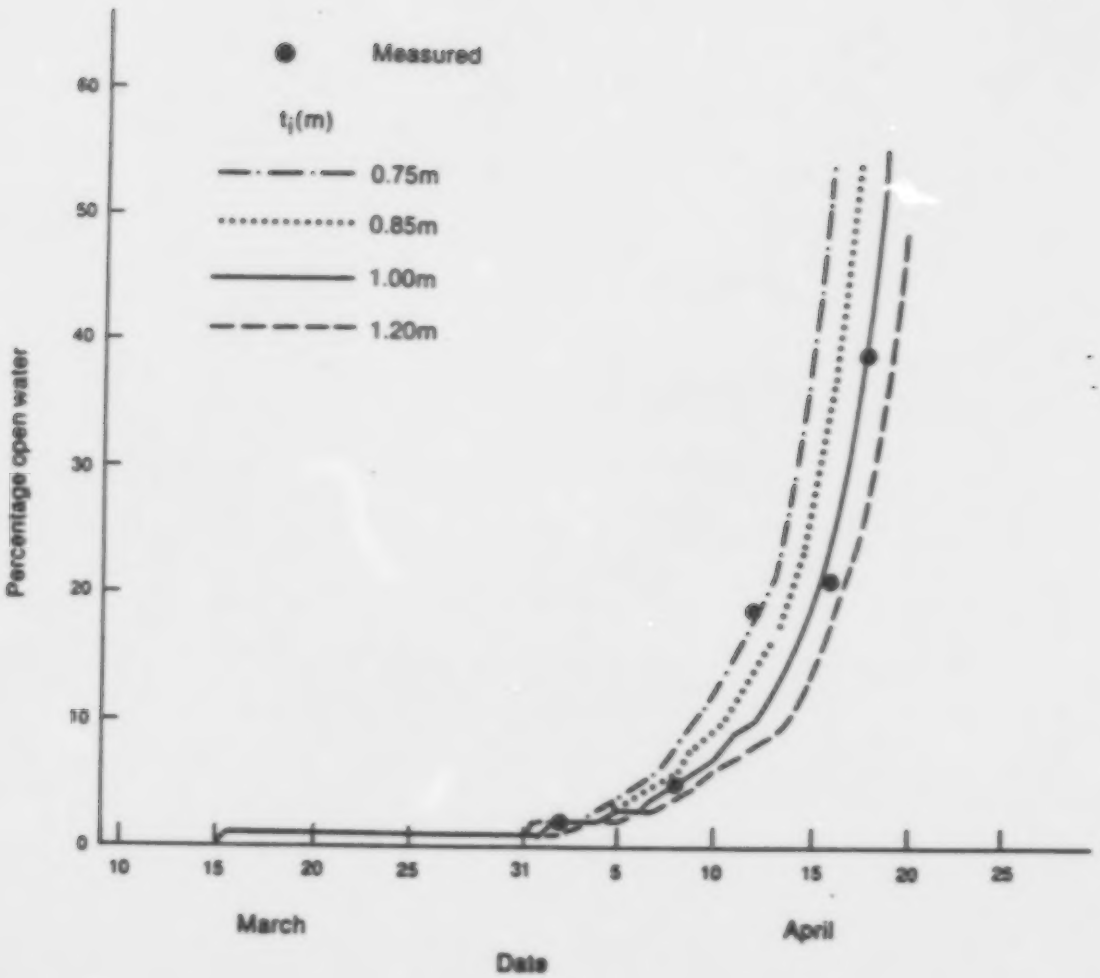


Figure 31 Sensitivity of the rate of open water generation to variations in the initial ice thickness.

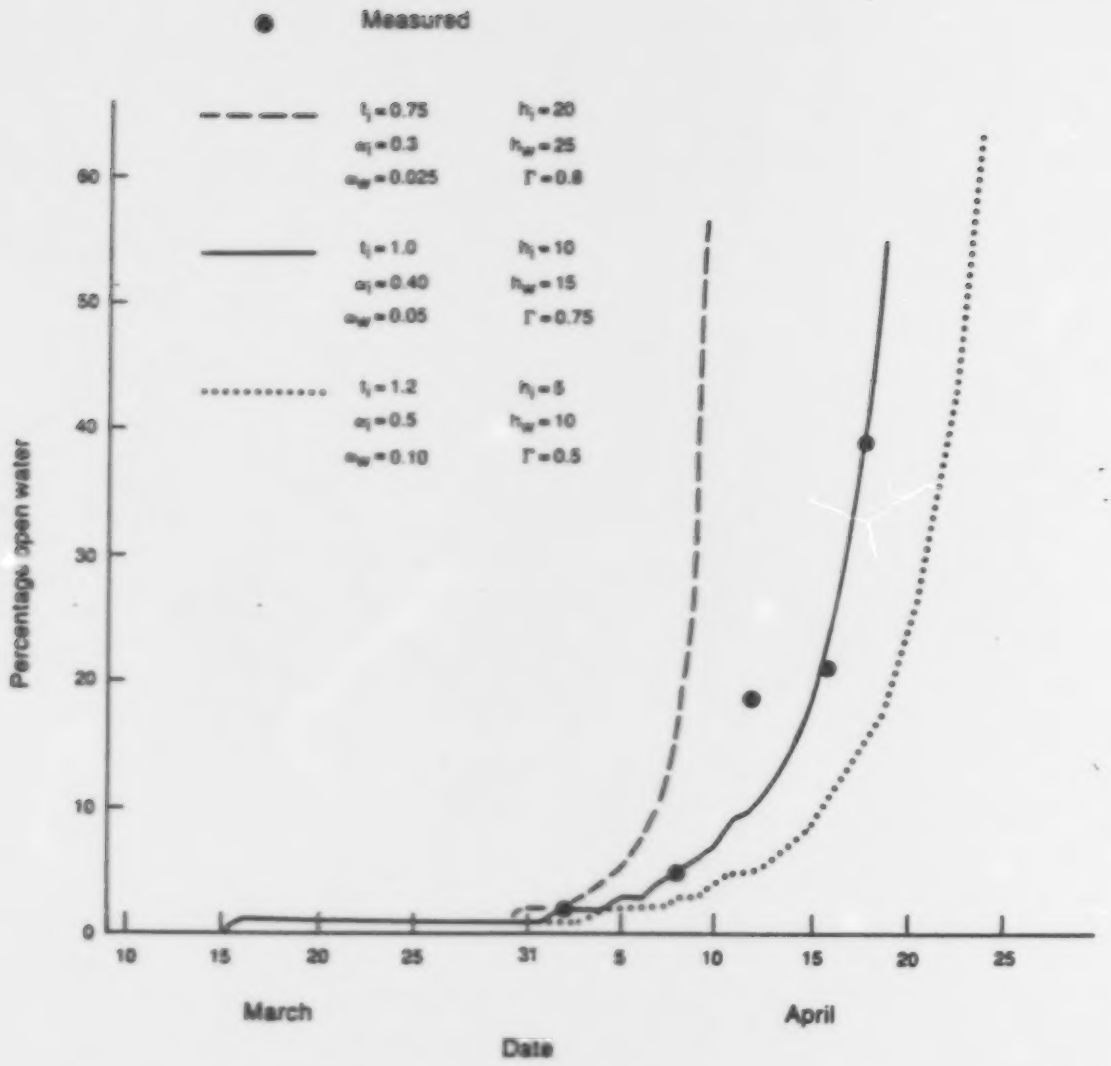


Figure 32 Upper and lower bounds of the open water generation rates for best and worst case scenarios of parameter estimates

that a window of some fourteen days can result using the poorest possible judgement of the model parameters.

The effects of the date of the initial stage increases prior to the appearance of open water on the rate of open water generated are shown in figure 33. In this situation, a stage increase prior to March 21, does not have an impact on the rate of progression towards breakup. Because the gauge height and discharge were constant before March 21, no stage increase was evident until March 21, and so there was no deviation from the calibration curve. However, when the first stage rise was set at time one day later, the reference width was subsequently increased and the measured stage increases after March 22 produced less initial open water causing a reduced fraction of open water available for solar radiation absorption and, therefore, delaying the onset of breakup significantly. Delaying the first day of stage increase by one day and by eight days would cause breakup to be delayed by two days and six days, respectively.

The effects of varying snowpack depths are summarized in figure 34. A relatively thick snowpack will take longer to melt, hence, delaying the beginning of ice thinning. On the other hand, if there is little snow, the snow cover will be depleted much sooner. The latter situation ultimately results in a thinner ice cover during the period of open water generation and, hence, produces breakup at an earlier date. For example, if the snowpack is totally removed (ice begins to thin) nine days earlier, breakup is advanced by two days. Even if the ice cover starts to thin three days earlier, there is an acceleration of breakup by about one day. Delaying the onset of ice cover thinning by ten days, delays breakup by only one to two days. Variations in this scenario significantly effect the ice thickness, which can vary between 0.3 and 0.8 m, depending on the day of initial thinning. Although the ice cover varies tremendously, the change in the date of breakup is not all that substantial.

This suggests that major process affecting the rate of progression towards breakup is the heat transfer across the air-water interface and

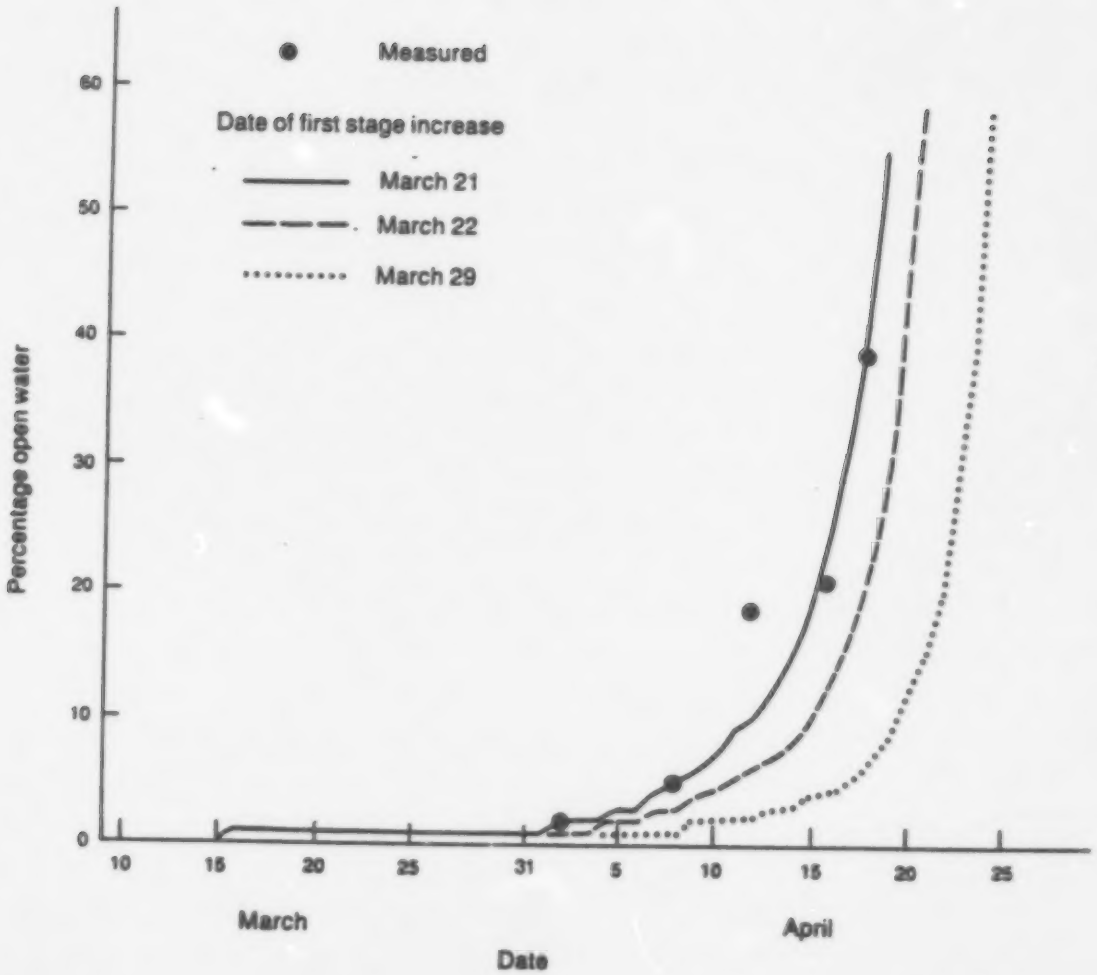


Figure 33 Sensitivity of the rate of open water generation to variations in the date of first stage increase

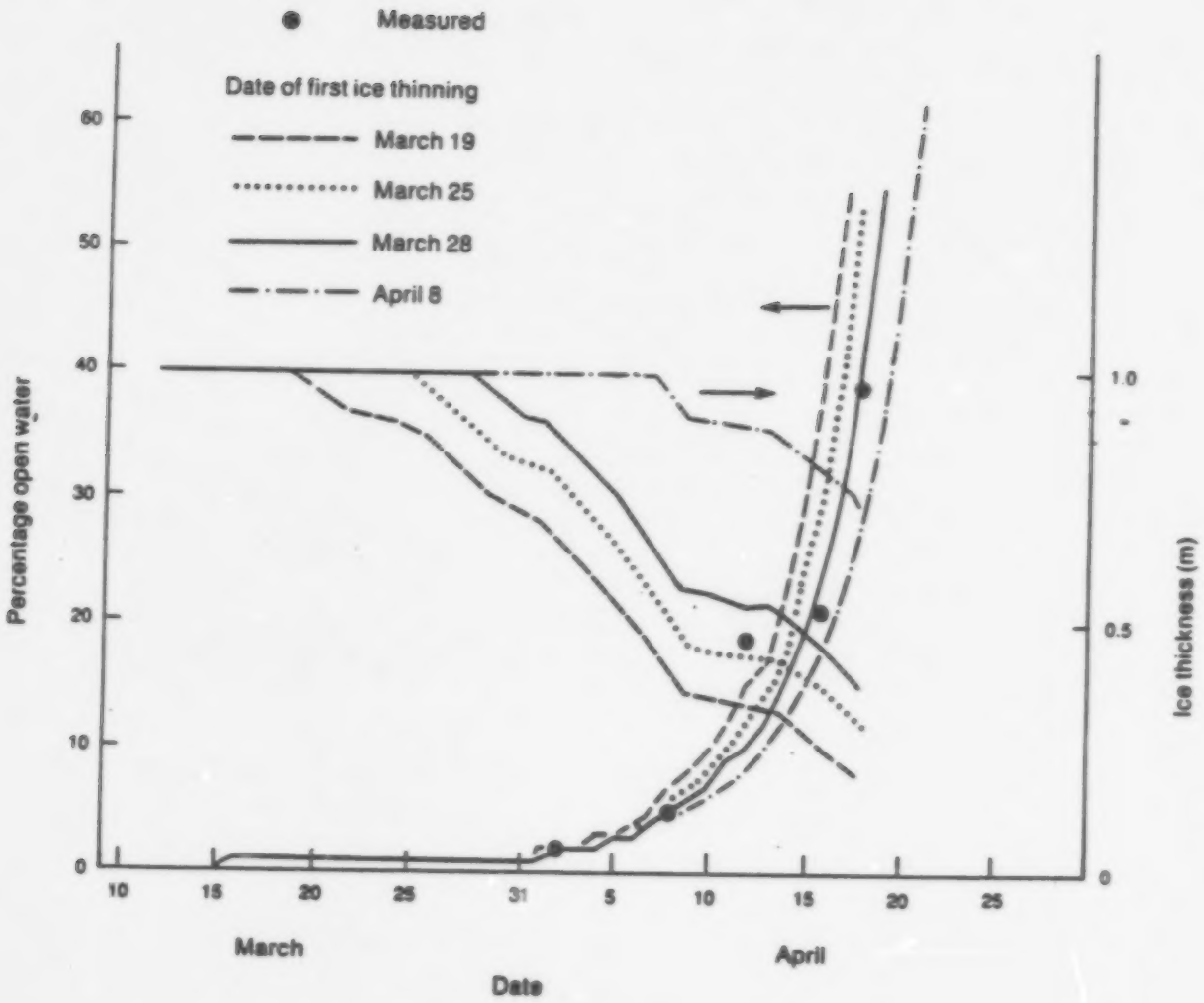


Figure 34 Sensitivity of the rate of open water generation and the ice thickness to variations in the date on which the ice cover begins to thin (date of last snow on the ice)

the process by which the warmer water melts out the intact ice cover. Nevertheless, for any given year, the breakup date is defined by a combination of characteristics related to the winter snowfall, ice thickness, snowmelt, runoff generation and the premelt (freeze-up) ice characteristics and river stage.

5.0 SUMMARY AND RECOMMENDATIONS

Generalized breakup over the entire reaction of the Athabasca River between the mouth of the Pembina River and Fort McMurray took place on April 19. This occurred after 62 °C-days of accumulated melt over a period of about 26 days. The net accumulated snowpack first began to be depleted on March 27 and was completely gone by the first week in April. Prior to breakup, the ice cover deterioration was characterized by a gradual increase in the fraction of open water from 0.02 on April 3 to 0.38 on April 18, the day prior to breakup. In general, the breakup could be characterized primarily as thermal and perhaps overmature, because very low stage increases (on the average) were apparent.

The observations of breakup summarized herein have led to the development of a model to predict the rate of open water development contributing towards incipient instability and breakup of a non-regulated ice covered stream. The observations indicate that for even slightly below normal snowfall, the breakup process is largely dominated by thermal processes and seems to be initially independent of the runoff. The initial open water is generated by a stage increase, resulting from the melting of the snowpack. This can occur before the snowpack is entirely removed and contributes to a decrease of the average albedo of the channel surface. This results in a more efficient transfer of energy (heat) across this surface and promotes the development of open water.

The ice cover begins to thin once the snow cover on the ice has been melted. The combination of a progressively thinner cover and a greater heat flux at the stream surface leads to an exponential increase in the fraction of open water. The processes by which this occurs are quite straightforward, and all can be rationalized quite easily except for the physics of the development of open water. This leads to the one crucial assumption in the model: all the heat that is absorbed by the fraction of open water goes to melting of a proportional amount of ice cover.

A numerical model, developed to account for these processes, was tested against one year of observations and was found to describe the

change in the fraction of open water relatively well. Using measured values of daily air temperature, solar radiation and discharge and initializing the model by defining the maximum winter ice thickness, the day of the first consistent increase and the last day of accumulated snow on the ice cover, it was found that the observations of open water could be calculated using an exposure factor of 0.75, an albedo of ice and water of 0.3 and 0.05, respectively, and the convective heat transfer coefficient over ice and water of 10 W/m^2-°C and 15 W/m^2-°C, respectively.

Sensitivity analysis suggested that the variations of the albedos and the heat transfer coefficients within appropriate and accepted limits did not significantly alter the model performance. It was most important to have a good definition of the maximum winter ice thickness, the date of first thinning (i.e., last day of a substantial snow cover) and the first day of appreciable rise in stage. For this specific location, it appears that an exposure factor of 0.75 is the most appropriate.

It should be noted that this model describes the process by which an ice cover goes from a very solid competent state to one so deteriorated that instability sets in and breakup occurs. For this specific year, the critical fraction of open water appeared to be about 0.40. However, in all likelihood, the critical value is a function of the increase in the stream flow. For years with a low snowpack and, hence, low flow increases, it is probably that Θ_c be relatively high. On the other hand, if there is a significant snowpack and considerable runoff develops, the fraction of open water needed to produce unstable conditions is probably less. Definition of the critical value, as function of runoff (or of the combination of snowpack and meteorologic conditions) can be done by looking at a number of years, applying the model to get a theoretical or computed Θ_c for the day prior to breakup and then for each of those years correlating Θ_c with the runoff or runoff potential.

Some other minor points also need additional clarification. The impact of cold spells on delaying the thermal melt have not been intensely rationalized. As a simple first order approximation, it was assumed that

freezing conditions doe not generate additional ice, rather, they simply delay the process until melting conditions are encountered once again. Another elementary, but, perhaps, important point is a better definition of the convective heat transfer coefficients, particularly as they apply to freezing and melting conditions. Also, some idea of the distribution of the snowpack on the ice cover and its melt rate (as compared to snow surveys in the basin) is required to better establish the date on which the ice cover begins to melt.

On the basis of the above discussion, given that the process by which incipient instability is reached has been described, it is recommended that further work be carried out to:

1. Define the critical fraction of open water, Θ_c for a number of years to determine its relationship to runoff potential. This can be achieved by applying the model for the years when the day of breakup can be defined, computing Θ_c at the breakup day and then relating it to the runoff potential.

2. Set up experiments to measure the rate of snow cover depletion and ice cover thinning, relate this to the basic meteorologic variables and define the importance of the albedo and the heat transfer coefficient.

3. Begin work to define the runoff on the basis of snowpack, antecedent moisture conditions and the meteorologic conclusions during the melt period. This should result in an enhanced ability to forecast snowmelt runoff and, therefore, improve the breakup prediction on the basis of the observed freeze-up and winter river stages, measured snowpack characteristics and the appropriate weather forecasts.

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